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1 Research Article

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

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11  
12 *Keywords*

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18 Erosion

19 Sedimentation

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20

21 **ABSTRACT**

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22 Sediment flux signals from source to sink in foreland basins preserve a record of tectonics, sea level  
23 and climate through erosion and sedimentation. However, longitudinal sediment transport often occurs  
24 in foreland basins, thus removing part of the orogenic material flux from foreland basin records. Here  
25 we use mass balance calculation and stratigraphic simulations of sediment fluxes for the Taiwan  
26 orogen to provide an order of magnitude estimate of how much orogenic material may bypass a  
27 foreland basin. Our results indicate a significant, potentially more than 50%, mismatch between  
28 sediment volume currently preserved in the basin and the amount of material presumably eroded from  
29 the orogen since the onset of collision in Taiwan. This suggests either a significant overestimation of  
30 average erosion rates over the period concerned with orogenic development of Taiwan, or it supports  
31 previous paleogeographic work suggesting that longitudinal sediment transport in the paleo-Taiwan  
32 Strait served as a major bypass conduit of importance for the establishment of a steady state orogen.  
33 We identify candidate submarine topography in the South China Sea that may preserve Taiwan's  
34 missing erosional mass.

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35

36 **1. Introduction**

37 Sediment fluxes within foreland basins exert a primary control on basin architecture involving  
38 interactions between tectonics, sea level and climate through erosion and sedimentation (e.g., Allen et  
39 al., 2013, Castellort et al., 2015, Flemings and Jordan, 1989, Posamentier and Allen, 1993). The  
40 orogenic history of many ancient basins has been reconstructed with help of sedimentary records, such

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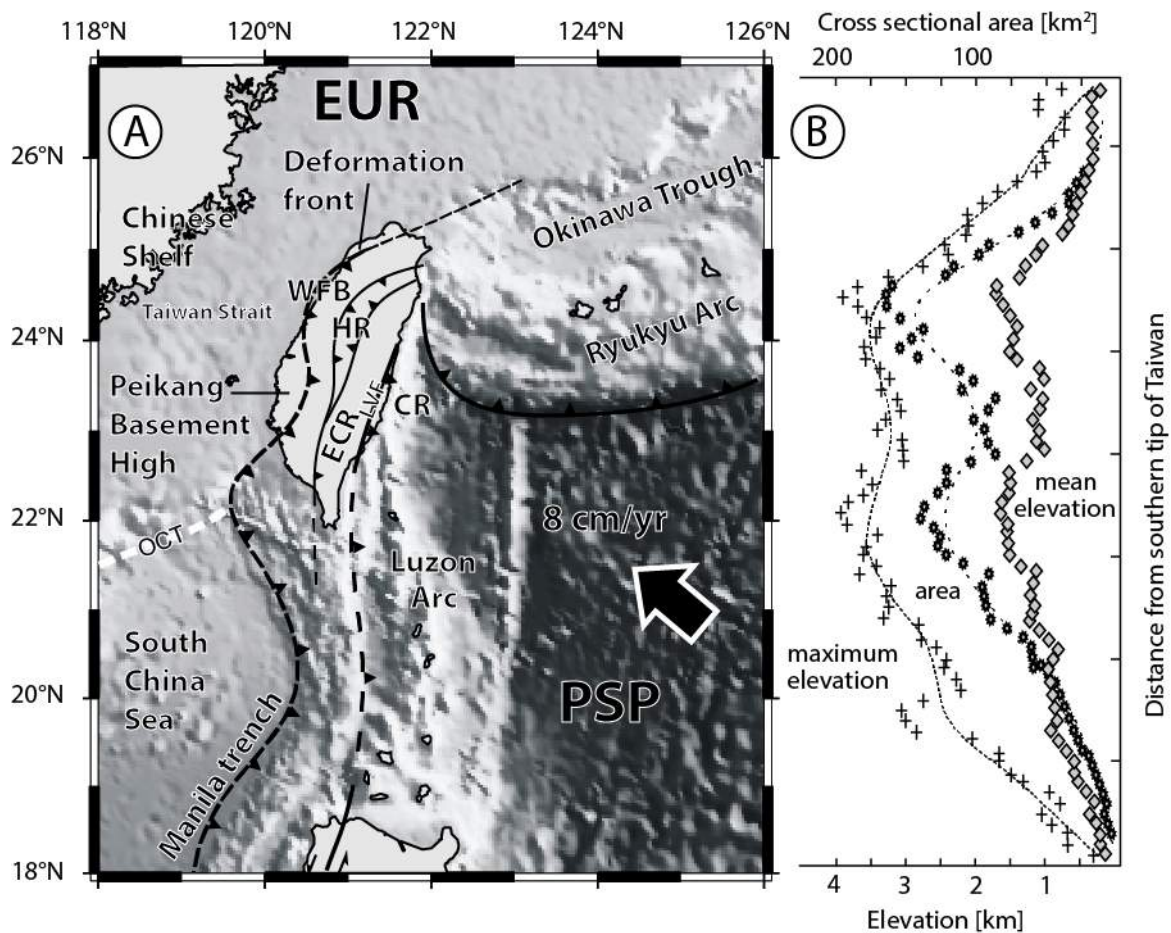
41 as in the Alps (Garzanti et al., 2004; Lihou and Allen, 1996), Pyrenees (Puigdefàbregas et al., 1992;  
42 Vergés and Burbank, 1996), or Himalayas (Garzanti et al., 2005; White et al., 2002), but it is still not  
43 well known how much of the orogenic history is eventually preserved and how tectonics, facies and  
44 sediment supply to basins are linked (Castelltort et al., 2015, Romans et al., 2016).

45 The western foreland basin in Taiwan (Fig. 1A) is a particularly suitable place to study interactions  
46 between tectonics and sediment fluxes because it is very young (5-6 Ma) and still very active (modern  
47 seismicity and extreme climate conditions). In this basin, southwestward ongoing oblique collision  
48 between the Luzon volcanic arc and the continental shelf of Eurasia (OCT: Ocean-Continent  
49 Transition on figure 1A) makes it possible to record the full evolution of basin deformation (e.g.,  
50 Suppe, 1981, Covey, 1984, Lin et al., 2003) and provides an opportunity to connect tectonics and  
51 depositional processes at different stages of the basin's evolution. The western foreland of Taiwan is  
52 the historical basin where the classical foreland filling sequence was first described by Covey (1984,  
53 1986). Indeed, the basin evolved from an early underfilled stage with relatively deep-water  
54 sedimentation (now observable in the modern setting in the South of the orogen) to a late balanced-  
55 filled stage, where shallow marine environments persist until today (most of modern Taiwan Strait,  
56 Covey, 1984), despite the enormous amount of sediment supplied to the ocean by the rising Taiwan  
57 mountains (Milliman and Kao, 2005; Milliman and Syvitski, 1992). In that sense, Taiwan orogen is  
58 emblematic of the distinct classical evolutionary stages (underfill to overfill, flysch to molasse) that  
59 characterize many ancient foreland basin systems such as in the Molasse basin of the Alps (Allen et  
60 al., 1991), the Bradanic Trough in the Apennines (Tropeano et al., 2002), the Solomon Sea in Papua  
61 New Guinea (Silver et al., 1991) or the South-Pyrenean foreland basin (Puigdefàbregas and Souquet,  
62 1986). As a consequence of oblique collision, the basin records a time-transgressive southwestward  
63 oriented migration of facies belts (e.g., Covey, 1984, Chen et al., 2001, Nagel et al., 2013) and  
64 sediment depocenters (Simoes and Avouac, 2006), similar to other oblique collisions such as in Papua  
65 New Guinea (Abbott et al., 1994; Silver et al., 1991), but the details of the geometry of the initial  
66 collision at the scale of Taiwan are still ambiguous and several models have been proposed. Whereas  
67 some models favor an arc-continent collision (Huang et al., 2006; Suppe, 1984, 1988; Teng, 1990),  
68 others suggested a two stage collision of an exotic block with the Eurasian continental margin and a  
69 second collision of Luzon volcanic arc with the passive margin (Lu and Hsü, 1992), or an arc-arc  
70 collision between Luzon volcanic arc and a paleo-Ryukyu arc system extending to the west of Taiwan  
71 (Seno and Kawanishi, 2009; Sibuet and Hsu, 1997; Sibuet et al., 1995), or even that collision may  
72 have happened synchronously along the margin at the scale of Taiwan (within a larger scale context of  
73 obliquity between EUR and PSP, Fig. 1A, Castelltort et al., 2011; Lee et al., 2015).

74 From an orogenic point of view, Taiwan has been proposed as a possible illustration of topographic  
75 steady state in a critical orogenic wedge because it shows an approximately constant width of 90 km  
76 (Suppe, 1981, Stolar et al., 2007) and elevations and cross-sectional area plateau in the central segment  
77 of the belt (Fig. 1B). Indeed, were it not at steady state, the orogen should be wider in the North were  
78 collision started earlier, and progressively narrower southwards. Instead the cylindrical shape of the  
79 orogen suggests that it has reached a critical size and slope. As a consequence, Taiwan orogen has  
80 been taken as an emblematic example of a steady state orogen in which erosional processes are able to  
81 balance uplift rates. As a note of caution, we remark that some authors have explained that the idea of  
82 steady state applies at large-scale in Taiwan but that, most probably, the high-frequency climate  
83 oscillations linked to orbital climate shifts, prevent the establishment of pure steady state at all scales  
84 (e.g., Whipple, 2001). Additionally, the Western foredeep itself also eventually reached a steady state  
85 size where accommodation space stayed constant despite the large sediment fluxes from Taiwan  
86 mountains (Covey, 1984; Covey, 1986). Therefore Covey (1986) suggested that sediment bypass out  
87 of the basin must have been an important factor that balanced accommodation space and sediment

88 supply, maintaining the basin shallow marine, and preventing it from becoming overfilled or even  
89 fully terrestrial.

90 The aim of this paper is to test, within the frame given by tectonic (plate boundaries configuration and  
91 timing, basin evolution) and geomorphic (steady state, erosion and sediment supply rates) constraints  
92 exposed above, the plausibility and magnitude of sediment bypass and to discuss implications for  
93 understanding foreland basins architecture within a well constrained source-to-sink setting. To do this,  
94 we use 3D stratigraphic simulations with different tectonic scenarios and we try to compare them with  
95 seismic lines from the Taiwan Strait to evaluate the match between simulations and observations.  
96 These simulations show how different tectonic settings control the stratigraphic evolution of the  
97 foreland basin, and allow to quantify sediment budget for the source-to-sink system. Results  
98 emphasize a significant mismatch between preserved volumes in the foreland with respect to volumes  
99 predicted given inferred erosion rates and topographic development. We discuss the implications of  
100 these contradictions for the importance of longitudinal sediment transport in foreland basins.  
101



102  
103 Figure 1. Geodynamic context of the collision in Taiwan and along-strike morphology. A) The plate context and main  
104 structural units in Taiwan. Note the overall obliquity between the Luzon Arc - Manila Trench system and the Eurasian  
105 margin (OCT). PSP: Philippine Sea Plate; EUR: Eurasian Plate; WF: western Foothills; HR: Hsüehshan Range; ECR:  
106 Eastern Central Range; LVF: Longitudinal Valley Fault; CR: Coastal Range; OCT: Ocean-Continent Transition. B)  
107 Longitudinal evolution of maximum elevation, mean elevation and cross-sectional area of the orogen (after Stolar et al.,  
108 2007, Suppe, 1981). The "plateau" of all three parameters displayed by the central segment of the orogen suggests that the  
109 mountain belt has reached a constant size and height despite the longer collisional history in the North compared to the  
110 South. Based on this assumption, Suppe (1981) among others have suggested that the orogen is at steady state, i.e. has  
111 reached a critical size for which tectonic influx of material is in overall balance with erosion. Decreasing elevation in the  
112 North is due to mountain range collapse associated with propagation of the extensional regime of the Okinawa Through. In  
113 the South, the orogen is still growing, has not reached steady state, hence the tapering elevation and decreasing width.

115

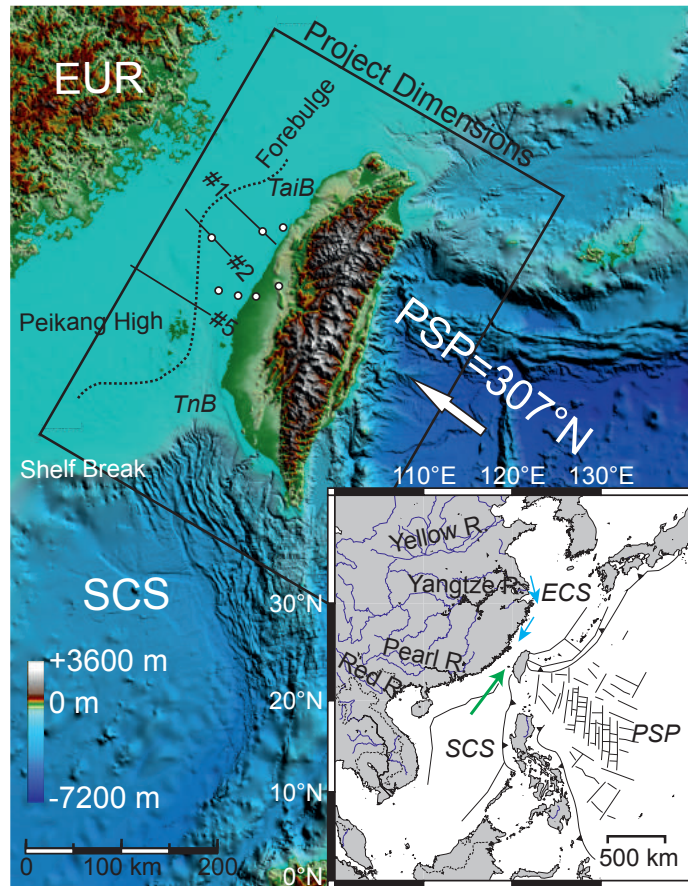
116 **2. General setting and background**117 *2.1. Geology and Tectonics*

118 The Taiwan mountains, rising almost 4 km above sea level, formed by collision between Philippine  
119 Sea plate and Eurasian continent shelf (Fig. 1A, Fig. 2). Arc volcanism associated with subduction  
120 below the Philippine sea plate ceased between 6 Ma and 3 Ma, when the arc resisted subduction and  
121 collided with the Asian passive margin to form an initial accretionary wedge (Huang et al., 2006;  
122 Yang et al., 1995). In most recent studies, arc-continent collision is estimated to have initiated in late  
123 Pliocene (Nagel et al., 2013; see discussion below). This is based on observing a continuous sandstone  
124 provenance shift and increasing illite crystallinity, interpreted to represent progressive unroofing and  
125 recycling of the metamorphic orogenic belt (Dorsey and Lundberg, 1988; Nagel et al., 2013). Oblique  
126 collision between the N-S trending Luzon volcanic arc and the NE-SW trending passive margin  
127 resulted in southwest propagating collision (e.g., Nagel et al., 2013; Simoes and Avouac, 2006; Suppe,  
128 1981; Teng, 1990), with modern collision point presently located offshore SW Taiwan (Lin et al.,  
129 2008; Yu and Huang, 2009). Today, the southernmost tip of Taiwan, which exhibits transient  
130 landscape features (Giletecz et al., 2015), represents the youngest relief associated with the emerging  
131 orogen. Oceanic lithosphere in the South China Sea is currently being subducted below the Philippine  
132 sea plate along the Manila Trench (Fig. 1A) whereas the Philippine sea plate itself is being subducted  
133 northwards below the Eurasian plate (Kao et al., 2000). This complex plate interaction manifests high  
134 active seismicity associated with a convergence rate of 70-80 km/Ma between the Philippine sea plate  
135 and the Eurasian continent (Seno et al., 1993; Wu et al., 2009; Wu et al., 2007; Yu et al., 1997). The  
136 current plate convergence is mainly accommodated within the Longitudinal Valley Fault (LVF, Fig.  
137 1A) on the east coast and at the deformation front in the Western Foothills consistent with the main  
138 active faults (Yu et al., 1997).

139 The continental margin experienced extensive rifting and continental breakup phases due to the  
140 opening of the South China Sea in late Paleogene, which resulted in major subsidence and numerous  
141 sub-basins separated by topographic highs (Lee and Lawver, 1995; Lin et al., 2003). This pre-  
142 collisional segmentation of the margin has an influence on the current structuration of the orogen into  
143 different tectonic units (Fig. 1A, Fig. 4B) that consist of (1) accreted volcanic arc (CR: Coastal Range)  
144 separated by suture zone (LVF: Longitudinal Valley Fault), (2) main orogenic belt (ECR: Eastern  
145 Central Range), (3) Hsuechan Range (HR), (4) deformed and uplifted foreland basin strata which  
146 constitutes a classical foreland with a fold-and-thrust belt (Western Foreland Basin), and (5)  
147 undeformed onshore (Coastal Plain) and offshore foreland basin sediments (Ho, 1988).

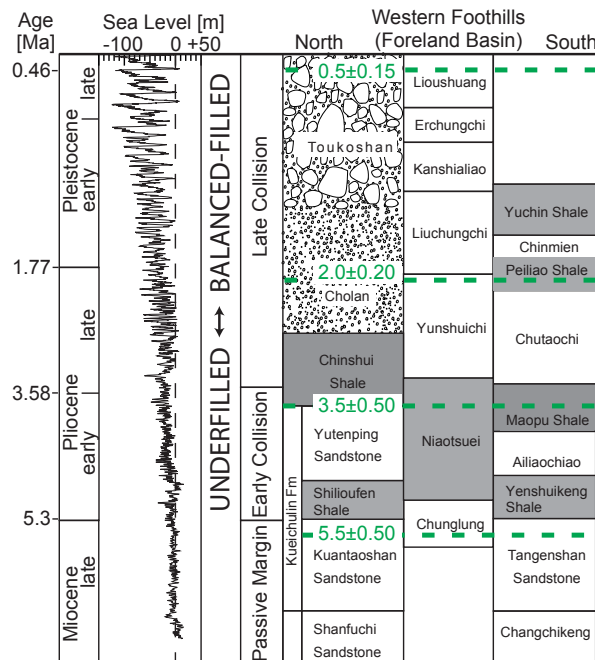
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149 **Figure 2:** Bathymetric and plate tectonic framework of the studied area showing boundaries of stratigraphic model. Inset  
 150 shows large-scale plate setting. Seismic lines tracks are from Yu and Chou (2001). Inset shows the four major Chinese river  
 151 systems in the area of Taiwan. Large green arrow is South China Sea Current (northward), and small blue arrows are China  
 152 Coastal Currents (southward). EUR: Eurasia. ECS: East China Sea. PSP: Philippines Sea Plate. SCS: South China Sea. TaiB:  
 153 Taishi Basin. TnB: Tainan Basin.  
 154  
 155

156 As initially described by Covey (1984), the evolution of syn-collisional facies is very similar from  
 157 north to south, except for distinctive grain size contrasts (Chou, 1973). The coarse fraction was  
 158 trapped in a shallow continental shelf basin (Taishi Basin, TaiB on Fig. 2), which was separated from  
 159 the South by the Peikang High topographic barrier (Fig. 1A and Fig. 2, Meng, 1967). Most of fine  
 160 grain sizes were transported further southwards and became deposited in a deep marine basin to the  
 161 South (Tainan Basin and/or South China Sea, TnB on Fig. 2).



162

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

170

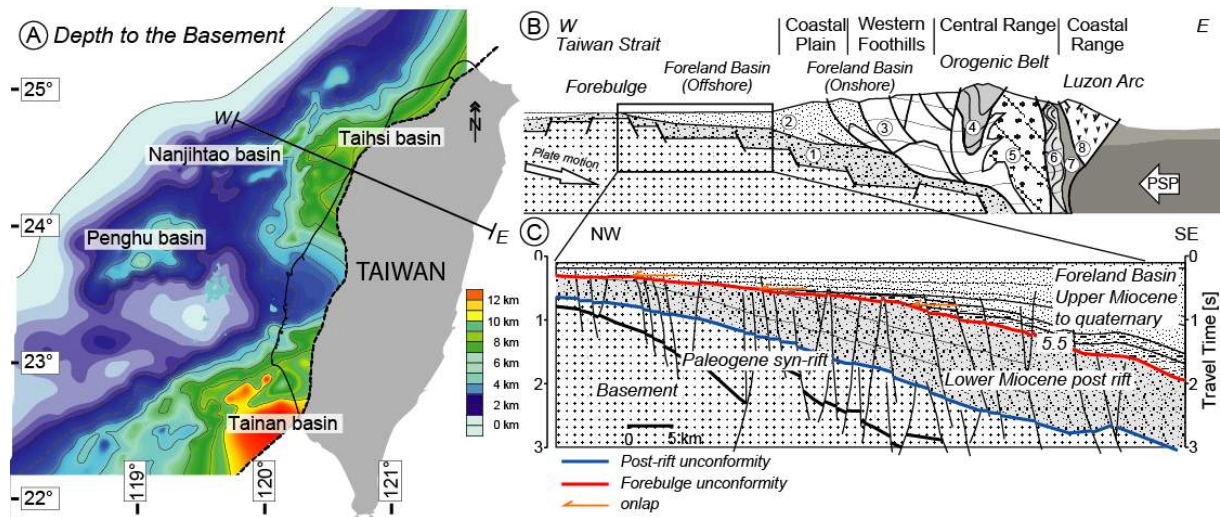
171 The chronostratigraphy of the Western Foothills has been extensively studied with Neogene  
172 calcareous nannofossils (Chang and Chi, 1983; Chou, 1973; Huang, 1977; Huang and Huang, 1984)  
173 and provides ground truth for five key biostratigraphic horizons (the most recent four are on figure 3)  
174 that are best documented (see synthesis in Nagel et al., 2013): nannofossil zone boundaries NN5-6  
175 ( $12.5\pm 1\text{Ma}$ ), NN11-12 ( $5.5\pm 0.5\text{Ma}$ ), NN15-16 ( $3.5\pm 0.5\text{Ma}$ ), NN18-19 ( $2\pm 0.2\text{Ma}$ ) and NN19-20  
176 ( $0.5\pm 0.15\text{Ma}$ ). The stratigraphic succession comprises a first retrogradational series consisting of  
177 shallow marine deltaic environments, which are often tidally influenced (Fig. 3, Kueichulin fm.). The  
178 transgression associated with the end of this formation marks the onset of orogenic loading of the  
179 shallow marine shelf environment. To the South, the formation passes progressively into deeper  
180 marine mud-dominated deposits (Fig. 3). The source of sediments during deposition of Kueichulin  
181 formation is essentially the same as during previous passive margin history of the basin, from the  
182 Eurasian continent to the southeast (Castelltort et al., 2011; Nagel et al., 2013; Shaw, 1996). It is  
183 followed by the Pliocene Chinshui Shale, a relatively deep marine mud-dominated formation, which  
184 represents the underfilled stage of the foreland basin (Covey, 1984). Reworked fossils, paleocurrent  
185 directions and facies analysis point to a main source from the East of the basin at this period, which is  
186 the growing orogenic wedge (Chang and Chi, 1983; Nagel, 2012; Nagel et al., 2013). The Cholan  
187 formation represents a large-scale progradational sequence of shallow marine wave- and tide-  
188 influenced environments, which became progressively dominated by fluvial processes upsection. This  
189 is the main foreland basin stage driven by large sediment fluxes out of Taiwan orogen and southward  
190 migration of facies belts. During late Pleistocene, increased erosion lead to deposition of large alluvial  
191 sediments, which most likely are an ancient example of braided rivers draining the orogen today  
192 (Covey, 1984).

193

## 194 2.2. The foreland basin unconformity

195 Flexural response due to loading of the Eurasian shelf by the forming orogen and its sedimentary  
196 response has been studied in detail (Castelltort et al., 2011; Chen et al., 2001a; Chiang et al., 2004;  
197 Simoes and Avouac, 2006; Tensi et al., 2006). Tensi et al. (2006) suggested that the passive margin  
198 lithosphere already experienced flexure since 12.5 Ma and interpreted the observed flexure as not  
199 being related to initial arc-continent collision, which is consistent with plate kinematic reconstructions  
200 (Hall, 1996; Nagel et al., in prep; Sibuet and Hsu, 2004). The basal foreland unconformity is observed  
201 in the Northern basins (Taishi basin, Fig. 4) with an age estimated between 8.6 and 5.6 Ma (based on  
202 biostratigraphic data), consistent with a flexural migration of the load from east to west (Lin and  
203 Watts, 2002; Lin et al., 2003). This unconformity separates the passive margin sequence and the  
204 foreland basin sequence, which onlaps onto it. The depositional hiatus increases in duration from the  
205 current frontal thrust towards the forebulge in the middle of Taiwan Strait (Lin et al., 2003; Yu and  
206 Chou, 2001).

207



208  
209  
210 **Figure 4:** A) Map of depth to the Cenozoic basement in Taiwan's foreland, highlighting four individual basins separated by  
211 basement highs. The Nanjihtao basin is partially covered by a distal foreland basin sequence. The Taihsi and Tainan basin are  
212 separated by the Peikang High. Figure modified from Lin and Watts (2003). B) Schematic cross-section of the Taiwan  
213 orogen in the North, where the interpreted forebulge topography is most pronounced. 1) Lower Miocene post-rift sequence,  
214 2) Upper Miocene to quaternary foreland sequence, 3) Miocene to Pleistocene sequences of the Western Foothills, 4) Eocene  
215 to Miocene metapelites, 5) and 6) Metamorphic formations of the Tananao Complex, 7) Neogenes sandstones and  
216 conglomerates, 8) Volcanic rocks and volcanoclastic of the Coastal Ranges (Luzon Arc). C) Detailed seismic line drawing  
217 modified from Yu and Chou (2001). The foreland basin sequence onlaps onto passive margin deposits, with increasing  
218 depositional hiatus towards the forebulge, where late Pleistocene-Quaternary sediments directly overlie lower Miocene strata.  
219

### 220 2.3. Modern sediment fluxes in Taiwan: uplift, climate and erosion

221 East Asian monsoonal climate was most probably established since 8.5 Ma (i.e., late Miocene) with an  
222 intensification observed since 5 to 3 Ma (Liu et al., 2003; Wan et al., 2006; Zheng et al., 2004). Today  
223 the island experiences 4 to 6 typhoons per year, with maximum mean annual rainfall of 2000-3000  
224 mm yr<sup>-1</sup> (Kao and Milliman, 2008). The total annual amount of sediment delivered to the ocean by  
225 Taiwanese rivers has been estimated to be up to 500 Mtyr<sup>-1</sup> with a strong asymmetry across the  
226 mountain range (Dadson et al., 2003; Liu et al., 2008). Estimates of erosion rates range between 2.2 to  
227 8.3 mmyr<sup>-1</sup> (Dadson et al., 2003; Fuller et al., 2003) and up to 30 mmyr<sup>-1</sup> (Resentini et al., 2017) in  
228 agreement with quantitative estimates from thermochronometric constraints of 3 to 10 mmyr<sup>-1</sup> (Lee et  
229 al., 2006; Willett et al., 2003).

230 Much of the suspended sediment is delivered at hyperpycnal concentrations into the Taiwan Strait  
231 (Dadson et al., 2005; Milliman and Kao, 2005; Milliman et al., 2007) where it is redistributed by  
232 seasonal and tidal currents (Jan et al., 2002). The northeast directed South China Sea current (green  
233 arrow, inset Fig. 2), for example, transports warm tropical water into the Strait with a peak intensity  
234 during summer months (June to August). In contrast, the southwest directed China Coastal current  
235 (blue arrows, Fig. 2) delivers Yangtze-derived mud into the northern Taiwan Strait (Hu et al., 2010;  
236 Xu et al., 2009) during winter months (September to May).

237 Marine observations indicate that fine mud particles are relatively quickly transported northwards out  
238 of the Taiwan Strait (Horng and Huh, 2011; Horng et al., 2012; Huh et al., 2011; Kao et al., 2008a;  
239 Liu et al., 2010; Milliman et al., 2007). For example, marine investigations in the Choshui river delta  
240 made before and after a typhoon hit the island, showed that fine-size particles are redistributed and  
241 transported northward within a month (Milliman et al., 2007). Thus, sediments eroded from the orogen  
242 possibly contained a larger amount of mud than currently found in ancient plio-pleistocene deposits,  
243 and that has been fractionated away by marine processes.



244 Average sedimentation rates vary greatly from 2 mm yr<sup>-1</sup> in the Western foreland basin to 3-4 mmyr<sup>-1</sup>  
 245 in the Coastal Range (Chen et al., 2001a; Lin et al., 2003; Lundberg and Dorsey, 1990), with a rapid  
 246 increase observed since the onset of deformation in the Western Foothills (Chang et al., 1983; Lock,  
 247 2007; Mouthereau and Lacombe, 2006; Mouthereau et al., 2001). These values are in accordance with  
 248 erosion rates estimates of between 2 and 10 mm yr<sup>-1</sup> from modern river sediment loads and  
 249 interpretation of thermochronological data (Dadson et al., 2003; Fuller et al., 2003; Fuller et al., 2006;  
 250 Liu et al., 2001; Liu et al., 2000; Siame et al., 2011; Simoes et al., 2007; Simoes and Avouac, 2006;  
 251 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 <sup>1</sup>	
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) <sup>2</sup>	
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 <sup>3</sup> 6 (CR), up to 60 <sup>4</sup> 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) <sup>5</sup>		4.2 (BR)-6.3 (TC) 2-3 (<1.5 Ma)
Fuller et al. (2006)	E: 3-5 (acceleration since 2-1 Ma)	<sup>1</sup> 2.3-3.3 (AFT/ZFT <sup>6</sup> ) 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) <sup>7</sup>
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)

<sup>1</sup>Hengchun Peninsula, Tainan area, Coastal Range

<sup>2</sup>Hengchun Peninsula, Coastal Range, Lanyu and Lutao

<sup>3</sup>SSC=calculation based on modern suspended sediment concentrations

<sup>4</sup>active thrust faults, Western Foothills, Southwest Taiwan

<sup>5</sup>thermokinematic modeling

<sup>6</sup>BR=Backbone Range, TC=Tananao complex, CR=Central Range, ECR=Eastern Central Range, CoR=Coastal Range

<sup>7</sup>AFT=Apatite fission track, ZFT=Zircon fission track

<sup>8</sup>Be10, Lanyang catchment

252

253

**Table 1:** Synthesis of observed and predicted uplift and erosion rate data for the Taiwan orogen.

254

### 255 3. Material and methods

256 As explained in introduction, the aim of this paper is to test the influence of different tectonomorphic  
 257 scenarios in controlling the stratigraphic architecture of the Western Foreland Basin of Taiwan, with a  
 258 focus on constraining the magnitude of sediment bypass out of the basin and its implications for  
 259 understanding the underfill to overfill evolution of foreland basins.

260 To do this we use a stratigraphic model that requires two main input conditions: subsidence/uplift and  
 261 sediment fluxes. In the following sections, we introduce the basic physical laws used in the 3D  
 262 stratigraphic model Dionisos and we then review the data and processing that are behind subsidence  
 263 and sediment flux patterns that we use in this work and that are based on observations reviewed in the  
 264 previous geological setting section. Finally, we outline the setup of numerical experiments.

### 265 3.1. 3D stratigraphic model "Dionisos"

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

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

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

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

274 where:

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

276  $Q_{water}$  = relative water flow [-]

277  $K$  = diffusion coefficient [ $m^2/yr$ ]

278  $h$  = ground elevation [m]

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

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

285 The study area was set as a 500 km x 320 km rectangle in the Taiwan Strait where abundant data is  
 286 available (Fig. 2). It is confined to the flexural forebulge in the West and to the Coastal Range in the  
 287 East, and includes the Taishi basin in the North and the Tainan Basin in the South (Fig. 2). The input  
 288 data required by Dionisos consist of tectonic subsidence for different time intervals, sediment supply  
 289 and eustatic sea level fluctuations. Parameters for compaction, flexure and sediment transport are  
 290 embedded within the model and can be adjusted to fit benchmark geological data. Sediment influx can  
 291 be set as a boundary condition into and/or out of the study area, but can also be simulated with  
 292 basement erosion.

293 The subsidence data is explained in the following section. Published boreholes and seismic lines  
 294 offshore in the Taiwan Strait and onshore (Lin and Watts, 2002; Lin et al., 2003; Yu and Chou, 2001),

295 together with constructed depth maps (Fig. 5) for five key stratigraphic horizons defined in an earlier  
 296 study (Nagel et al., 2013) provide a solid first approximation database. Constraints on sediment fluxes  
 297 are obtained from combining modern river loads, denudation data and sediment volumes and  
 298 presented in sub-section 3.3.

299

### 300 3.2. Foreland basin subsidence

301 The most important basin-scale controls on accommodation include sea-level changes and flexural  
 302 subsidence related to lithospheric thickening and tectonic loads. The West Taiwan basin formed by  
 303 flexural bending of the Asian passive margin in front of westward migrating thrust loads of the  
 304 growing accretionary wedge (Lin et al., 2003). In order to better constrain subsidence of the  
 305 sedimentary basin, backstripping techniques (using Airy isostasy) were applied to 28 boreholes and 9  
 306 stratigraphic sections (Nagel et al., 2013; Watts and Ryan, 1976).

307 Backstripping is used to stepwise decompact and unload a borehole or stratigraphic section. The  
 308 influence of water depth and sedimentation on total subsidence are extracted in order to isolate the  
 309 contribution of tectonic subsidence alone. The tectonically driven subsidence at any location in the  
 310 basin is given in Allen and Allen (2009):

$$TS = Y \cdot \left( \frac{\rho_m - \rho_s}{\rho_m - \rho_w} \right) - \Delta sl \cdot \left( \frac{\rho_w}{\rho_m - \rho_w} \right) + (W_d - \Delta sl)$$

311 where  $W_d$  is the average water depth at which the sedimentary units were deposited,  $Y$  is the  
 312 decompact sediment thickness, and  $\rho_m$ ,  $\rho_w$ , and  $\rho_s$  are densities of mantle, water and sediment, and  
 313  $\Delta sl$  is the difference in sea-level height  $h$  between the present and the time at which the sediments  
 314 were deposited:

$$\Delta sl = \left( \frac{\rho_m - \rho_w}{\rho_m} \right) \cdot (h_2 - h_1)$$

315 The water depth at the time of deposition for the backstripped strata was estimated using  
 316 paleobathymetries from depositional models of Nagel et al. (2013). Note that since sediments in the  
 317 western foreland basin were deposited on a shallow marine continental shelf, the influence of the  
 318 water column (10s of metres) on the backstripped strata is small relative to the considered sediment  
 319 thicknesses (100s of metres). For decompaction, sediment was assumed to be composed of two main  
 320 grain size classes, sand and mud, which correspond to the modern siliciclastic river supply and is  
 321 consistent with detailed lithologic analysis (Huh et al., 2011; Nagel et al., 2013). When the basin gets  
 322 progressively filled with sediments, mechanical compaction introduces loss of water during sediment  
 323 burial and affects depth-porosity curves for different lithologies. The trend between porosity and depth  
 324 is usually approximated by:

$$\phi = \phi_0 \cdot e^{-cy}$$

325 This produces an asymptotically low porosity with increasing depth, where  $\phi_0$  describes surface  
 326 porosity and  $c$  the coefficient of compaction (Table 2). The flexure of the basement was computed  
 327 with an elastic thickness of 15 km, a Young's modulus of 100 GPa and a Poisson's ratio of 0.25,  
 328 chosen according to published values for the Taiwan foreland basin (Lin and Watts, 2002).

329

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

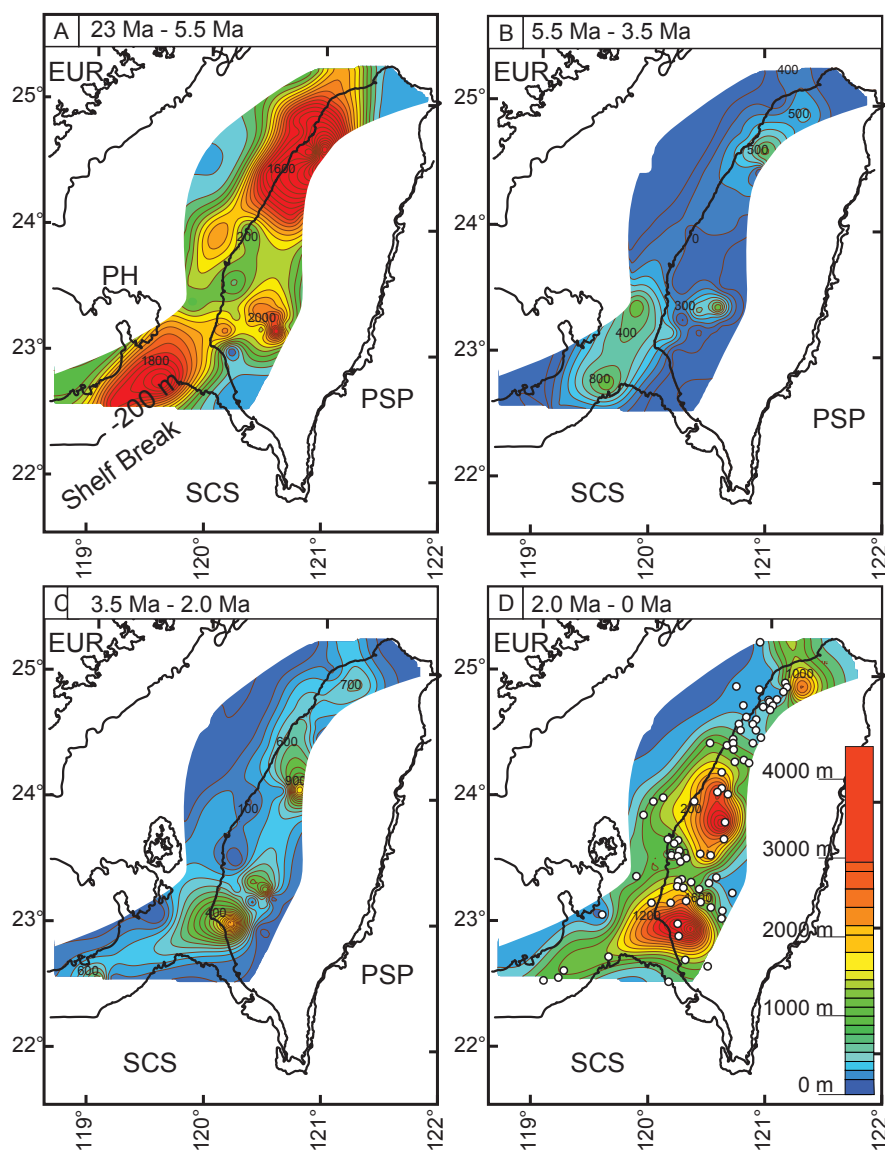
330

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

333

334 Backstripping results provide a detailed record of the Asian passive margin subsidence and uplift  
 335 history at the five key biostratigraphic horizons explained above (Fig. 5). Lin *et al* (2003) showed that  
 336 the subsidence history of the Asian passive margin is strongly influenced by its syn- and post-rift  
 337 history due to extension in the South China Sea (post-breakup extension from 30 to 21 Ma, thermal  
 338 subsidence from 21 to 12.5 Ma and a second post-breakup extension from 12.5 to 6.5 Ma). Increased  
 339 subsidence since the early Pliocene (Fig. 5D and Fig. 6) is ascribed to growth of the Taiwan orogen as  
 340 it propagates westward, introducing deformation and increasing sedimentation rates in the basin  
 341 (Chang and Chi, 1983; Mouthereau et al., 2001). In addition, Tensi et al. (2006) demonstrated that the  
 342 load associated with initial foreland basin has migrated rapidly westward 1 Ma ago and was stabilized  
 343 at the same time as the basin was buried under large quantities of sediments (alluvial and fluvial fans  
 344 of the Toukoshan fm.).

345



346

347 **Figure 5:** Maps of decompacted sediment thickness in between the five key biostratigraphic (nannofossils) horizons of Nagel  
 348 et al (2013). White dots in map of panel D represent location of Chinese Petroleum Company drill holes.

349

350 The reconstructed subsidence pattern (Fig. 6) is consistent with sediment isopach maps shown in  
 351 Figure 5. The obtained subsidence history of each stratigraphic section can then be interpolated to the  
 352 boundaries of the stratigraphic simulation project to produce subsidence maps (Fig. 7) that constitute a  
 353 direct input into Dionisos. Subsidence pattern in between two key stratigraphic horizons is simply  
 354 linearly interpreted between the two considered maps.

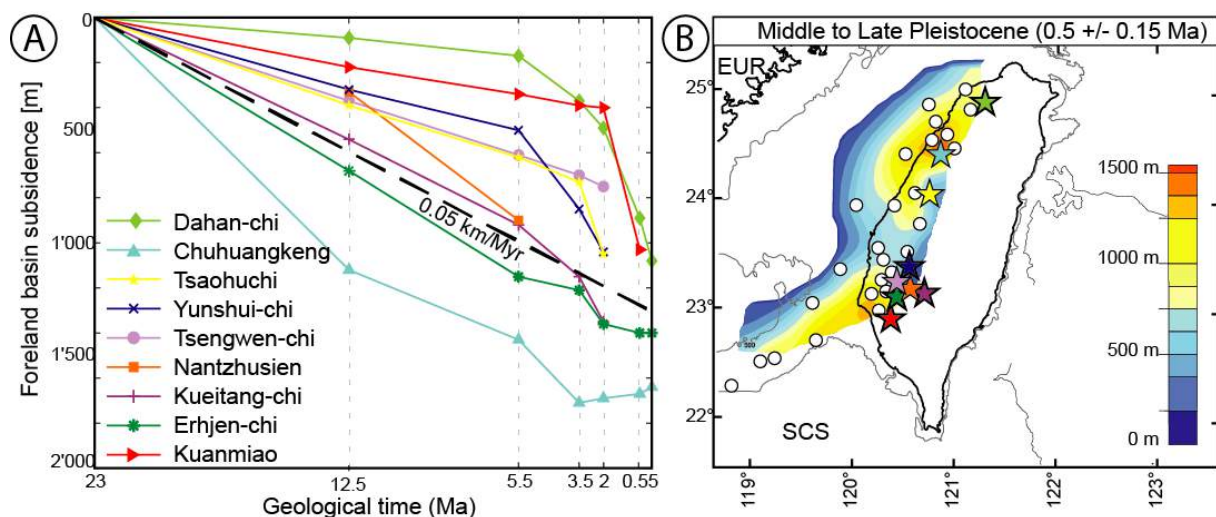
355 Lin and Watts (2002) showed that topography is insufficiently high to produce the observed  
 356 subsidence pattern in an isostatic flexural model driven by surface loads. Following Simpson (2014), it  
 357 can be proposed that this observation is an illustration of a possible decoupling between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

362

### 363 3.3. Sediment fluxes and basin boundaries

364 The volume of sediment deposited in the basin was calculated from data of published boreholes drilled  
 365 by the CPC (Chinese Petroleum Corporation) in the western foreland basin (Fig. 5) (e.g., Lin et al.,  
 366 2003; Shaw, 1996). For each sequence the sediment thickness is extrapolated between the present day  
 367 forebulge and the Western Foothills by a triangulation algorithm to obtain four maps between early  
 368 Miocene and late Pleistocene (four maps in between the five key chronostratigraphic horizons, Fig. 5).

369

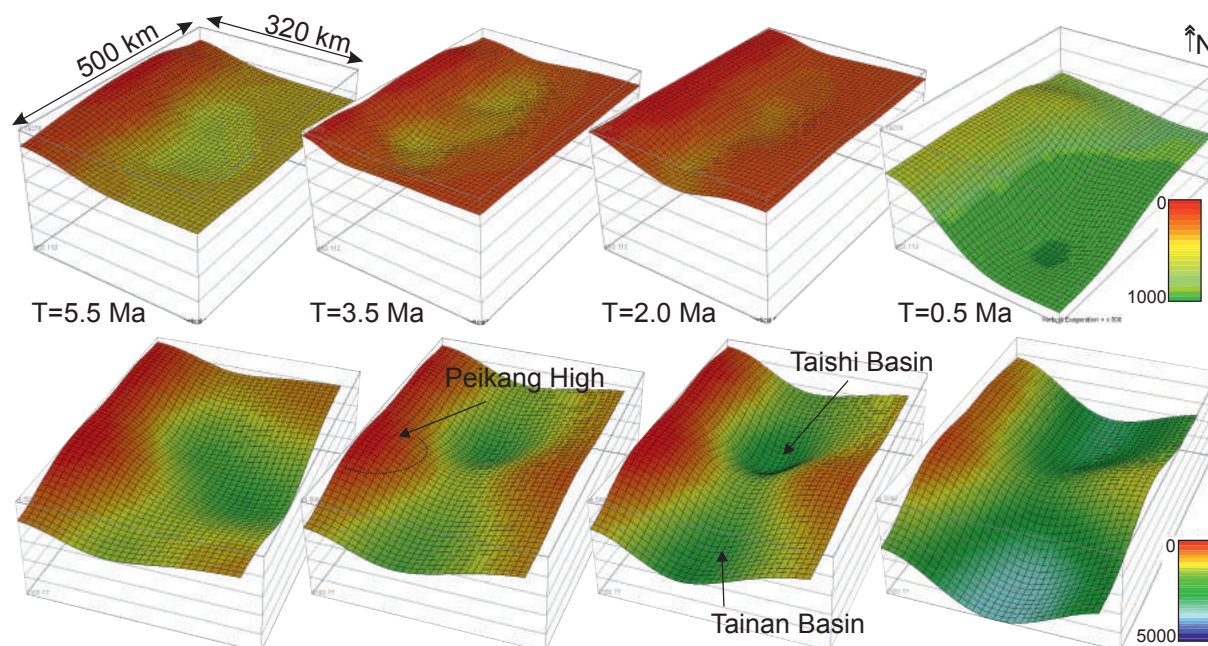


370

371 **Figure 6:** A) Tectonic subsidence histories from stratigraphic sections (each colored line correspond to a section) in the fold-  
 372 and-thrust belt of Taiwan (marked with a star, ordered from north to south). The vertical dashed lines correspond to the five  
 373 key chronostratigraphic horizons (see text) chosen to subdivide the foreland basin evolution in 4 stages. Dahan-chi (Pan,  
 374 2011), Chuhuangkeng (Huang, 1976), Yunshuichi/Tsaohuchi (Yeh and Chang, 1991; Yeh and Yang, 1994), Tsengwen-chi  
 375 (Chen et al., 2001a), Nantzhusien (Ting et al., 1991; Yu et al., 2008), Kueitangchi (Huang, 1977), Erhjen-chi (Hornig and  
 376 Shea, 1994), Kuanmiao (Chiu, 1975). B) Tectonic subsidence map for the late Pleistocene (NN19/20) with 28 boreholes and  
 377 9 stratigraphic sections. Stars indicate stratigraphic sections with color coding corresponding with left panel in which sections  
 378 legend is ordered from North (top) to South (bottom).

379





380  
 381 **Figure 7:** Smoothed subsidence maps for each simulated interval with tectonic subsidence used in the program (upper row)  
 382 and total subsidence (including isostatic effects of sediment and water loading, lower row). The sediment depocenters in the  
 383 north (e.g. Taishi basin) and the south (e.g. Tainan basin) are separated by the Peikang High.

384  
 385

386 The sediment volume accumulated within each time sequence is shown in Table 3. A total of 82'000 to  
 387 125'000 km<sup>3</sup> of sediment accumulated since 5.5 Ma in the foreland basin of Taiwan. If we assume that

Age [Ma]	Sediment Volume [km <sup>3</sup> ]	Sediment Flux [km <sup>3</sup> /Ma]	Sediment Flux [km <sup>3</sup> /Ma] From Asia	Sediment Flux [km <sup>3</sup> /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
<i>Total</i>	<i>67'657 - 100'864</i>			
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
<i>Total</i>	<i>159'872 - 230'749</i>			

388 collision started between 5.5 Ma and 3.5 Ma, and that before 5.5 Ma accumulated sediment thickness  
 389 corresponded primarily to influx of material from Asia mainland, we interpret the increasing sediment  
 390 influx from the Taiwan orogenic wedge to be in a range of 6'500 to 28'000 km<sup>3</sup>/Ma (Table 3). This  
 391 sediment flux is probably overestimated since the basin also must have received material from its  
 392 western border, i.e. Asia mainland. However, this contribution was probably swamped by the dramatic  
 393 increase in sedimentation rates that accompanied Taiwan orogeny (Chang and Chi, 1983). In addition,  
 394 some of the sediment transported into the foredeep consists of recycled foreland basin deposits.  
 395 Therefore, calculated sedimentation rates over the area of modern foreland basin are lower when  
 396 compared with sedimentation rates from Western Foothills, especially during the last phase of  
 397 orogenesis from 2 Ma to 0 Ma (Chang and Chi, 1983).

398

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

402

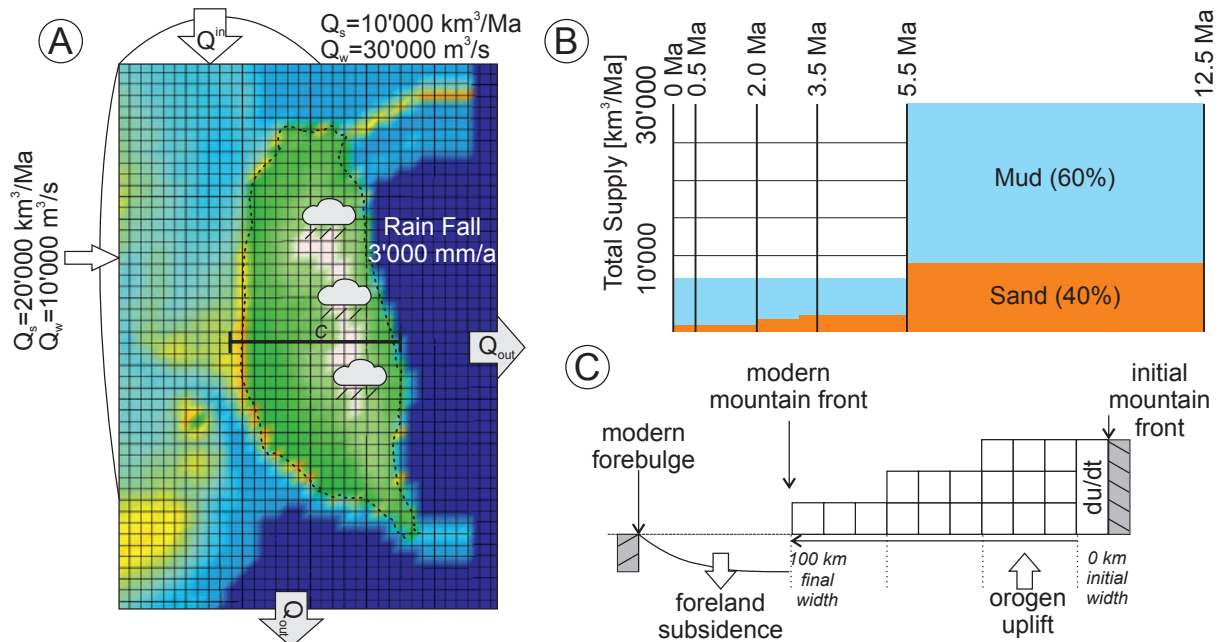
403 **Table 4:** Sediment volume accumulated in the Cenozoic sedimentary basins of Southeast Asia (modified from, Métivier et  
 404 al., 1999).

405

406 In addition to previous consideration, sediment fluxes from the growing Taiwan orogen are  
 407 constrained by comparing the amount of sediment that has been preserved in Taiwan foreland basin

Sediment Volumes SE Asia [km <sup>3</sup> ]		2 - 0 Ma	2 - 5 Ma	5 - 11 Ma	11 - 17 Ma	17 - 24 Ma	24 - 5 Ma	[km <sup>3</sup> /Ma]
Pearl River & S. Taiwan	Total	81000	93000	240000	85000	130000	455000	23947
	uncertainty	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
	uncertainty	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
	uncertainty	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

408 with the amount of sediment accumulated in Cenozoic sedimentary basins north and south of Taiwan  
 409 (Table 4). The obtained boundary supply fluxes are minimum, because some unknown amount might  
 410 have bypassed or not even reached the Taiwan Strait.



411

412

413 **Figure 8:** Initial sediment supply history applied in the model. A) The two source areas correspond to Asia mainland (West)  
 414 and the East China Sea (North). Rainfall follows the modern mean annual rainfall rate. The fluvial water discharge was  
 415 estimated by modern river discharges in Southeast Asia (Table 5). B) The initial total sediment supply from the boundaries  
 416 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

417 measurements in the Taiwan Strait (Huh et al., 2011; Kao and Milliman, 2008; Xu et al., 2009), as well as observations in the  
 418 ancient sedimentary record (Nagel et al., 2013). See text for discussion. C) Sketch of orogen growth as implemented in  
 419 Dionisos: the area of uplift rate is progressively enlarged to simulate mountain range widening, from 0 at the onset to 100 km  
 420 width at the end.

421  
 422 Finally, initial sediment supply history imposed in the model is shown in Figure 8. Two sources of  
 423 sediment are defined along the western and northern sides of the model box (Fig. 2). It is important to  
 424 note that these sources refer to general provenances located along the model boundaries and are not  
 425 meant to represent individual rivers. Today, only smaller tributaries of the Minjiang and Jiulong rivers  
 426 drain directly into the Taiwan Strait, collectively discharging only 1/10 of Taiwanese rivers (Table 5).  
 427 Water discharge per source area was assumed to be similar to modern water discharge of rivers in  
 428 Southeast Asia (Table 5).

	catchment area [ $km^2$ ]	mean channel gradient [m/km]	average water dis- charge [ $m^3/s$ ]	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	78000	-	8000	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

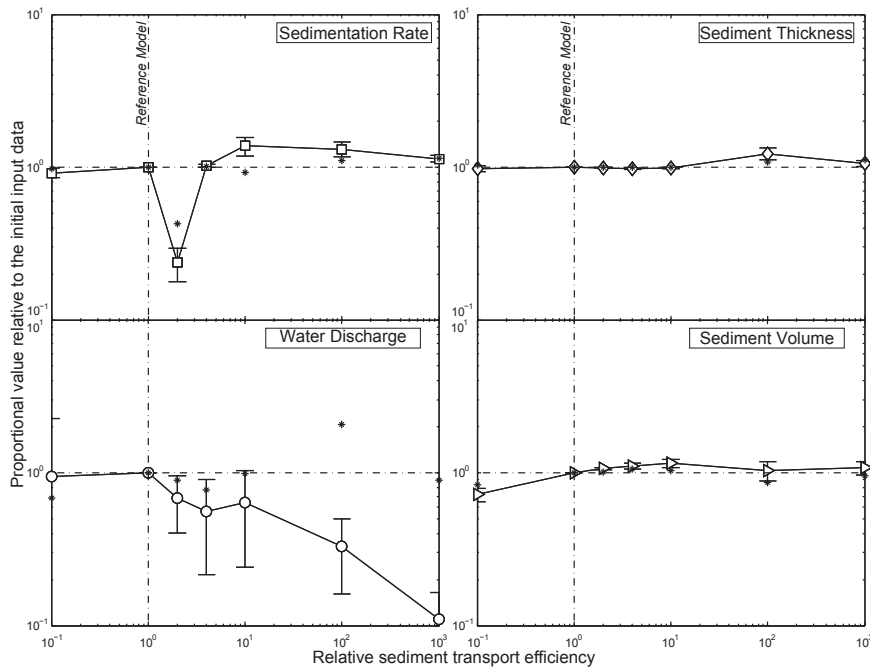
429  
 430 **Table 5:** Main parameters of rivers from Taiwan, Southeast Asia and larger rivers worldwide for comparison. Data World  
 431 from Somme *et al* (2009) and references therein. Data for Southeast Asia from Chen *et al* (2001b), Hicks *et al* (2000), Huh  
 432 *et al* (2011), Kineke *et al* (2000), Kniskern *et al* (2010), Kuehl *et al* (2004), Liu *et al* (2007), Milliman and Syvitski (1992),  
 433 Olariu and Steel (2009), Wang *et al* (2007), Wang *et al* (2005), Wolanski *et al* (1995), Yu *et al* (1991), Yu and Huang (2006)  
 434 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*  
 435 (1997).

436  
 437 Since sediment transport in Dionisos is modeled by diffusion, a short review of published values for  
 438 the diffusivity coefficient K in different depositional environments is provided here for the sake of  
 439 comparison with other studies having used a similar approach (Table 6). Although these studies did  
 440 not all use diffusion in exactly the same way for modeling sediment transport (for instance depending  
 441 on whether water discharge is taken into consideration or not), an average value for each depositional  
 442 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	

443  
444 **Table 6** Ranges of diffusion coefficients used in modeling studies for individual depositional environments. The values were  
445 converted to [km<sup>2</sup>/ka].

446  
447 Figure 9 shows the sensitivity of different parameters (water discharge, sediment thickness, sediment  
448 volume, sedimentation rate) for 7 model runs with increasing sediment transport efficiencies (between  
449 0.1xK<sub>initial</sub> and 1000xK<sub>initial</sub>, with K<sub>initial</sub> being the diffusion coefficient of base model). All models  
450 were run with the standard model set-up described above. Parameters were measured at three different  
451 points within the basin at seismic lines #1, #2, and #5 (see Fig. 2) as well as the average value from  
452 3.5 Ma to 0 Ma (marked with asterisk).

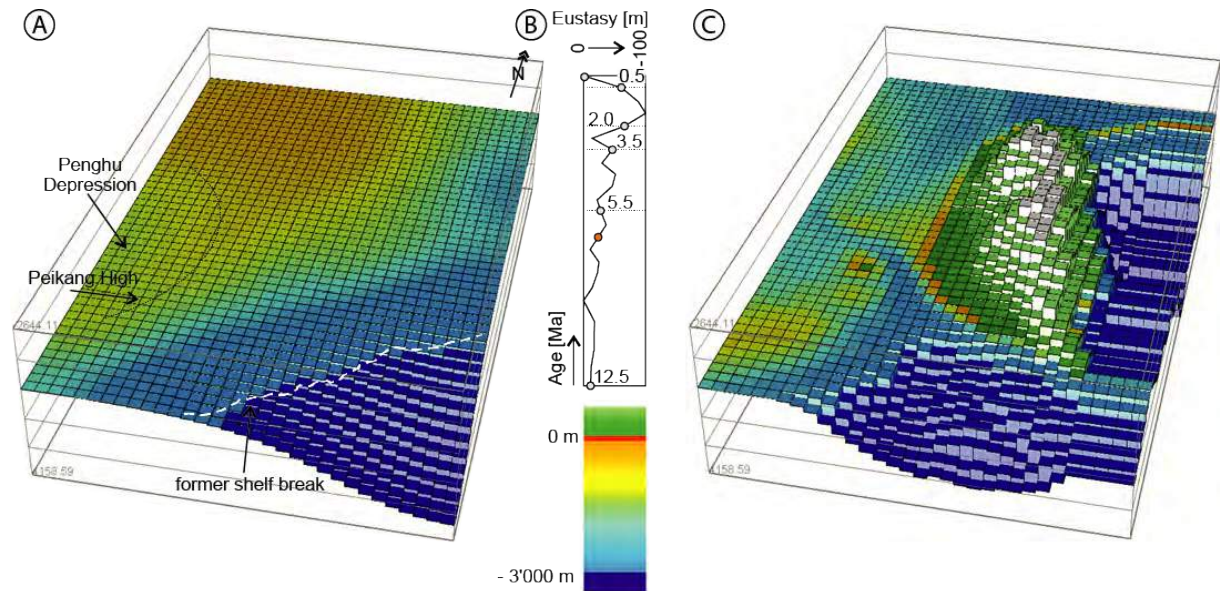


453  
454 **Figure 9:** Sensitivity test for different diffusion coefficients. Proportional values of the four measured parameters in relation  
455 to the initial model input data within the foreland basin, plotted against the relative efficiency of sediment transport. The  
456 results are from seven separate model runs with the same initial reference model parameters, but different diffusion  
457 coefficients.

458  
459 Simulations start at 12.5 Ma, which corresponds to the NN5-6 nannofossil boundary. This key  
460 biostratigraphic horizon has already been used in an earlier study to reconstruct the paleogeography  
461 during the arc-continent collision (Nagel et al., 2013). The study shows that sedimentation in the  
462 foreland basin during the Miocene to Pleistocene took place in a mixed storm- and tide-dominated



463 shallow marine depositional environment. Paleobathymetry did not change significantly from 12.5 Ma  
 464 to 3.5 Ma (Fig. 10), when the basin started to subside due to the approaching orogenic wedge in the  
 465 east and the mud-dominated Chinshui Shale was deposited (Fig. 3). It is important to note that  
 466 progradation and shallowing-upward cycles associated with the approaching orogenic wedge took  
 467 place earlier in the northern parts of the basin and progressed southward as the basin was filling up  
 468 (Nagel et al., 2013).



469

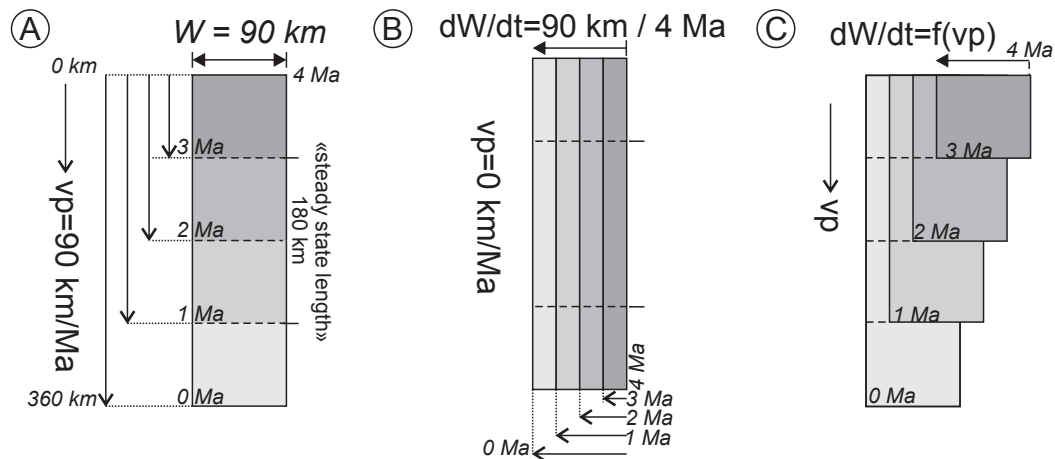
470 **Figure 10:** Initial bathymetry map at 12.5 Ma (A), and modern bathymetry (C). The 12.5 Ma map was constructed based on  
 471 detailed facies analysis in the Western Foothills (Nagel et al., 2013). Sea level variation (B) after Miller et al. (2011) for  
 472 indication.

473

#### 474 3.4. Experimental setup

475 To explore the orogen growth history and basin architecture, three different tectonic scenarios were  
 476 tested (Fig. 11). Each model considers the same initial boundary supply data (Fig. 8). In these  
 477 experiments, orogenic uplift begins at  $\sim 4$  Ma, which is in agreement with recent provenance studies  
 478 (Nagel et al., 2013). The first orogenic growth model (Fig. 11A) considers southward propagation of  
 479 the orogen at a rate of 90 km/Ma until present day length of 360 km is reached, and assumes a fixed  
 480 steady state width of 90 km (Suppe, 1981). Using the time-space principle initially constructed by  
 481 Suppe (1981), steady state size was reached after *ca* 1.3 Ma following onset of orogeny. In a second  
 482 model (Fig. 11B) it is assumed that the orogen collided with a large promontory simultaneously along  
 483 the length of the modern orogen, with no (or just minor) southward propagation. This scenario is based  
 484 on sedimentological studies and paleogeographic reconstruction of Castellort et al (2011) and  
 485 tectonic-thermochronometric data of Lee et al (2015). The third model, intermediate between both  
 486 previous ones, (Fig. 11C) considers a linear growth in length of the orogen with time, along with  
 487 lateral displacement of the orogen area as it overthrusts the Eurasian margin. In all three models, a  
 488 continuous and constant uplift rate of 5 km/Ma was assumed. This rate covers the range of uplift rates  
 489 that have been determined in Taiwan (Table 1).



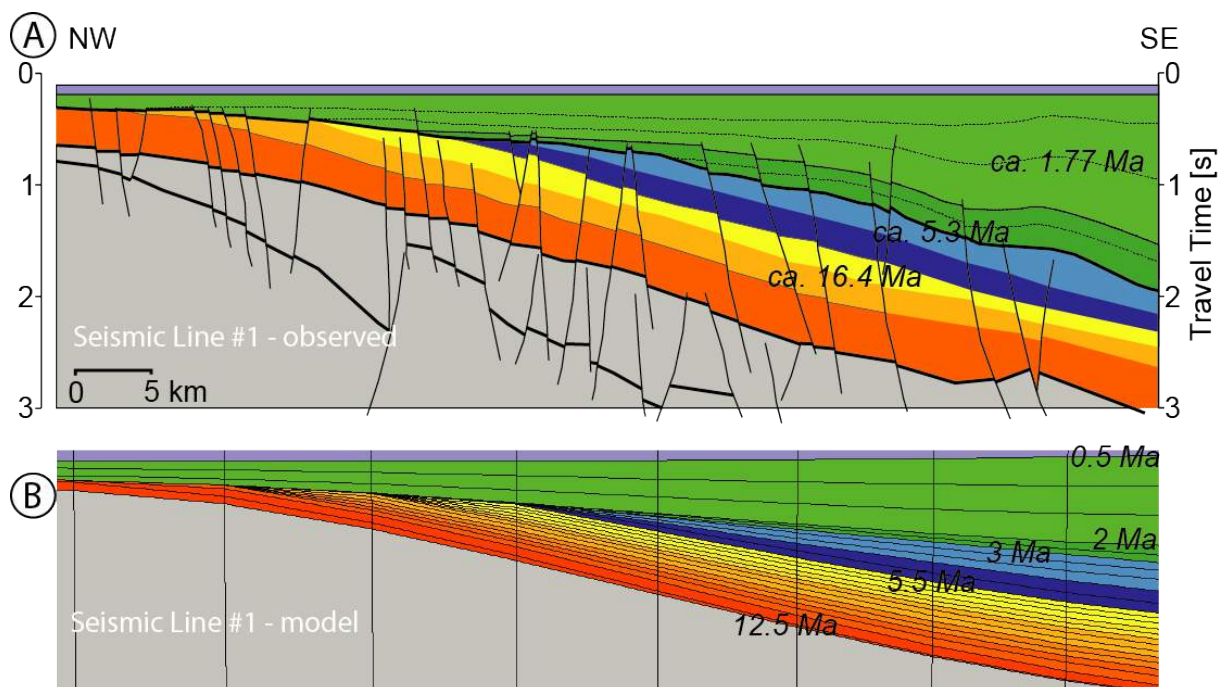


490  
 491 **Figure 11:** Three orogen growth models tested in this study. A) Pure lengthening: southward propagation (90 km/Ma) of a  
 492 steady state orogen with a fixed width of 90 km. B) Pure widening: lateral propagation, with a fixed length of 360 km. C)  
 493 Lengthening and overthrusting: southwestward propagation of a steady state orogen.  
 494

#### 495 4. Simulation results and discussion

##### 496 4.1. Foreland basin geometry

497 An initial test was performed to explore the adequacy of the imposed basin subsidence to reproduce  
 498 the first order geometry observed on seismic lines in the Taiwan Strait (Fig. 4). A key horizon to  
 499 compare is the transition from passive margin sedimentation to foreland basin sedimentation with the  
 500 flexural forebulge unconformity as described above (Lin et al., 2003). As shown in Figure 12, the  
 501 imposed timing and subsidence results in stratigraphic patterns (Fig. 12B) that correlate well with  
 502 geometries observed in seismics (Fig. 12A).  
 503



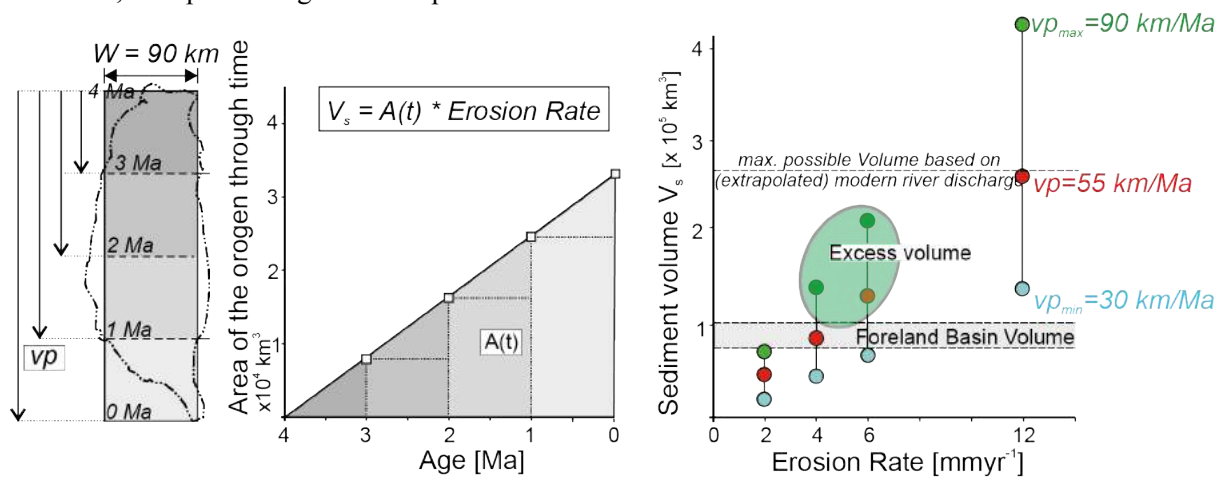
504  
 505 **Figure 12:** A) Yu and Chou interpreted seismic line #1 (see location on Fig. 2). B) Cross-section at the position of  
 506 seismic line #1 through simulated stratigraphy with Dionisos (this study). The input subsidence forces a dramatic change  
 507 of sedimentation pattern at the transition from passive margin sedimentation to foreland sedimentation. This mimics the  
 508 "flexural forebulge unconformity" documented by Yu and Chou (2001). This unconformity represents the boundary between

509 the pre-collisional Nanchuang Fm. and the syn-collisional Kueichulin Fm. and was estimated approximately at 6.5 Ma (Lin et  
 510 al., 2003).

511

#### 512 4.1. Mass flux calculations

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



523

524 **Figure 13:** A) Orogen growth model with a steady state orogen width of 90 km and a southward propagation rate of  $V_p=30$   
 525 to 90 km/Ma. B) The theoretical volume eroded from the mountain was calculated by integrating the orogen area through  
 526 time multiplied by the erosion rate. C) The modern river discharge was extrapolated over 3.5 Ma and taken as an upper limit  
 527 for the maximum possible sediment influx into the basin system (Table 5). The theoretical sediment volume eroded from the  
 528 mountains and feeding the Western Foreland Basin was corrected for the fluvial discharge flowing to the east, which is  
 529 currently 45% (Dadson et al., 2003).

530

531 To explore sediment dynamics within the foreland basin, mass balance calculations were done for a  
 532 southward propagating orogen model. The total amount of material transported into the basin  
 533 (according to tectonic scenario of Fig. 13) is compared with the amount of sediment preserved. The  
 534 theoretical total amount of material, which has been eroded from the orogen since 4 Ma, is estimated  
 535 by multiplying the integrated area (Fig. 13) with erosion rate (Table 1). Currently 55% of the annual  
 536 fluvial sediment discharge is flowing West and 45% is drained East (Dadson et al., 2003; Liu et al.,  
 537 2008). Hence the total sediment volume produced by the orogen was corrected for fluvial discharge  
 538 flowing East.

539 Figure 13 shows the potential sediment flux into the foredeep coming from the Taiwan mountains as  
 540 estimated by the model. Computed fluxes vary from 25'000 km<sup>3</sup> (for a southward propagation rate of  
 541 30 km/Ma and an average erosion rate of 2 mm/yr) and up to 425'000 km<sup>3</sup> (for a southward  
 542 propagation rate of 90 km/Ma and an average erosion rate of 12 mm/yr), although this may be  
 543 overestimated since it does not take into account the recycling of foredeep sediments. The current river  
 544 sediment flux during typhoon season (Liu et al., 2008) was taken as an upper boundary for the  
 545 maximum possible sediment influx, when extrapolated over 4 Ma (i.e., 285'000 km<sup>3</sup>).

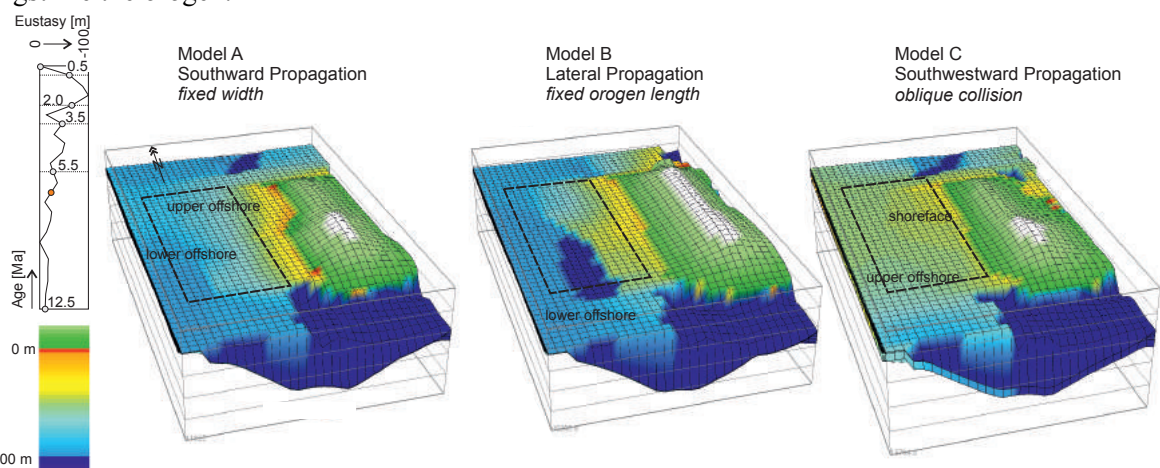
546 The amount of sediment preserved in the basin is much smaller than the possible amount of sediment  
 547 brought into the basin (Fig. 13) when one considers a southward propagation rate of 90 km/Ma and  
 548 erosion rates in a range of 4 to 6 km/Ma (Table 1). Our calculations suggest that between 25'000 km<sup>3</sup>  
 549 and 115'000 km<sup>3</sup> of material may have bypassed the foreland basin. If this is correct, it suggests that at  
 550 least half the sediment eroded from the orogen may not be preserved in eventual foreland basin  
 551 stratigraphic records. This material is likely longitudinally transported South, out of the basin (Nagel  
 552 et al., 2013), and into the South China Sea. Observations in south-central Taiwan already indicate  
 553 enhanced southward sediment transport since Late Pliocene marked by increasing amounts of  
 554 submarine incisions (Fuh et al., 2003; Fuh et al., 1997). The southward sediment transport is also  
 555 observed in the migration of sediment depocenters and facies belts, mainly driven by large sediment  
 556 flux from the orogen (Nagel et al., 2013; Simoes et al., 2007).

557

#### 558 4.2. Simulated sediment fluxes

559 An orogen that produces a steady flux of sediments was modeled for each of the three different growth  
 560 scenarios in Figure 10 and the volume of material deposited in the basin was calculated for each  
 561 scenario (Fig. 14). Steady state is established when the elevation of the mountain's top reaches a  
 562 roughly constant value in less than 1 Ma. This is achieved by tuning the diffusion coefficient for  
 563 continental sediment transport  $K$ , whereby an increase in  $K$  equals an increase in erosion, until a value  
 564 of  $K$  is found that gives satisfactory results (mountain range elevation) for all 3 scenarios. Three  
 565 different models were run, with a mean uplift rate set to 3, 5 and 12 mm/yr. Material is allowed to  
 566 leave the basin to the south by diffusion. The area of the orogen at each time step is the same for each  
 567 growth model, thus with identical uplift and erosion parameters the available material at each interval  
 568 is assumed to be equal. This allows to compare the three models only in terms of their different  
 569 tectonic growth scenario, and in terms of their implications for sediment transport in the basin  
 570 alongstrike the orogen.

571



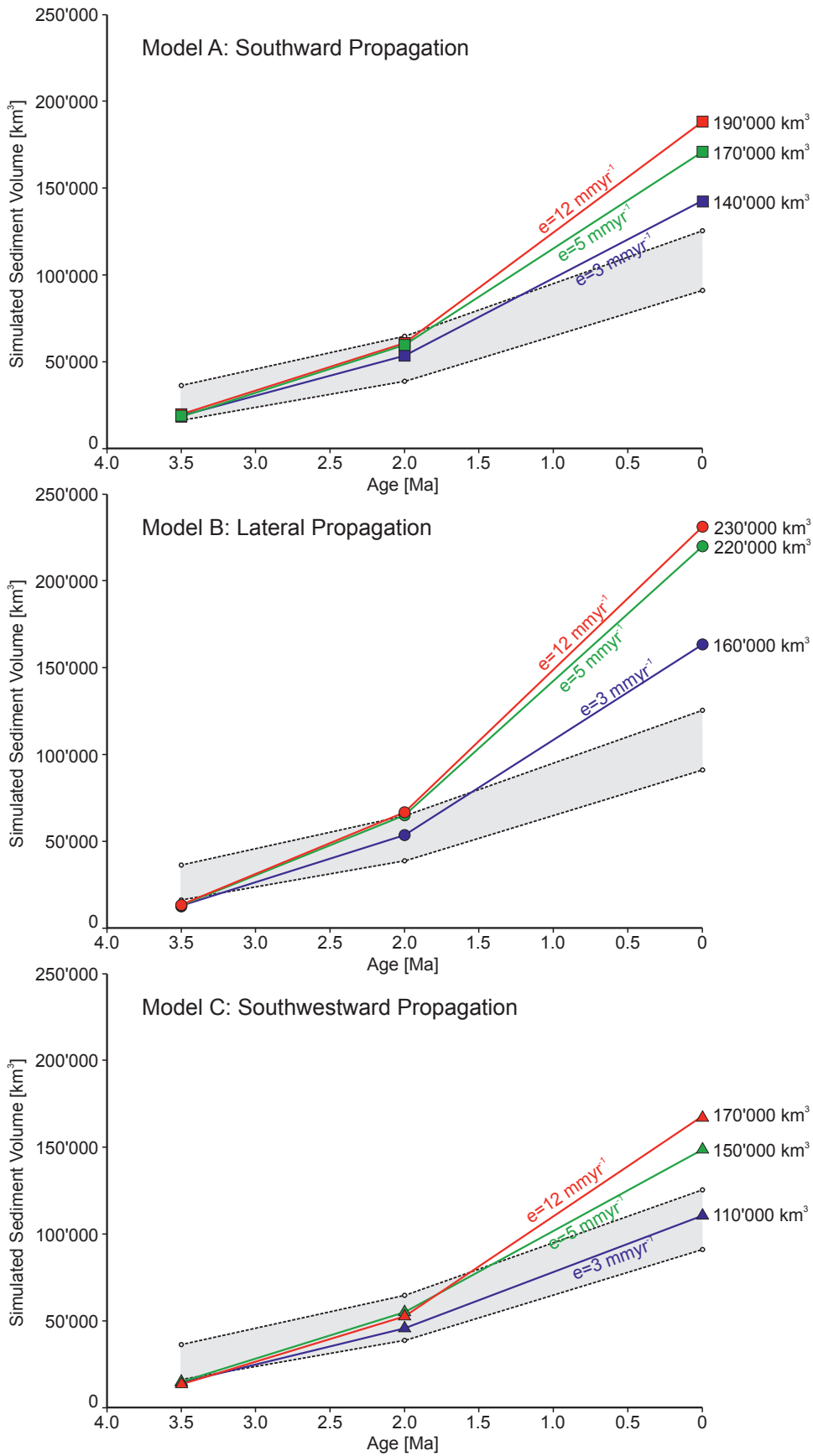
572

573 **Figure 14:** The model setup is shown schematically in Fig. 10. Model A is simulated with a southward propagation rate of 90  
 574 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  
 575 combined " oblique" collision, or southwestward propagation. In all the three models, the final orogen area and final  
 576 erosional fluxes vary only by minor amount due to different erosional landscape evolution during relief growth. The area  
 577 where the simulated foreland basin volume was measured is indicated with a black box.

578

579 The sediment volume of the foreland basin produced by each of the three models is shown in Figure  
 580 15. The three standard models (southward, lateral, or oblique propagation) tend to overestimate the  
 581 preserved sediment volume. Southward and oblique propagation achieve a better fit to observed  
 sediment thicknesses than lateral propagation. Moreover lateral propagation did not accurately

582 reproduce foreland basin geometries. The best fit (geometry and volume) is achieved with the oblique  
 583 collision scenario.



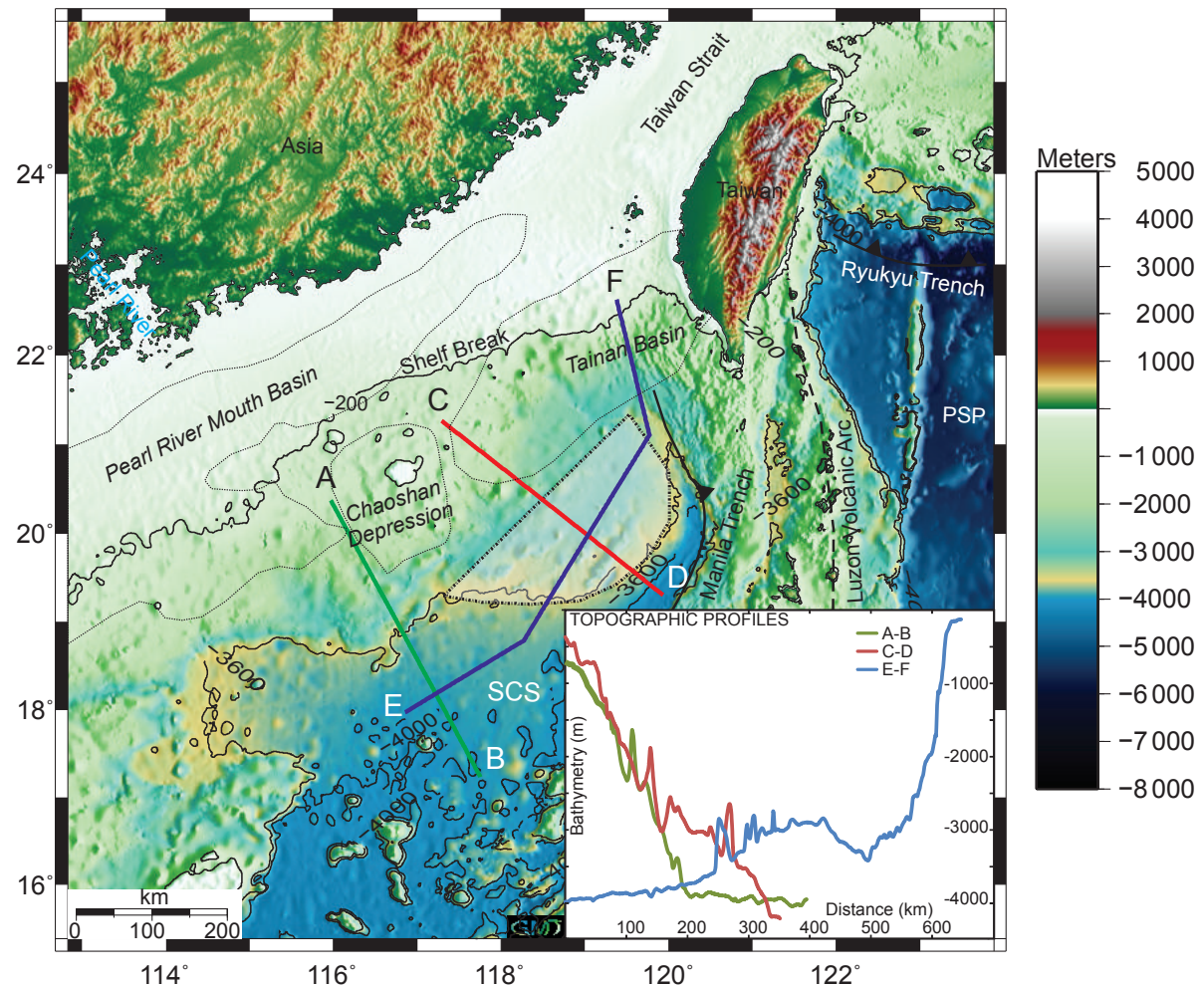
585 **Figure 15:** Calculated sediment volume in the foreland basin produced with the three standard collision models (Fig. 10).  
586 The preserved sediment thickness in the Taiwan foreland basin is between 70-125'000 km<sup>3</sup> (shaded area, see also Table 3).

587  
588 The southward propagation models suggest an excess of sediment carried into the basin of between  
589 15'000 and 80'000 km<sup>3</sup>. This amount is in agreement with the theoretical mass balance calculations  
590 (Fig. 13). As observed, even though the orogen reached a steady state size as suggested by Suppe  
591 (1981), due to longitudinal transport, the basin never becomes overfilled.

592 Earlier observations already implied an important longitudinal sediment transport out of the basin and  
593 observations from the southwest of Taiwan seem to confirm these predictions (Covey, 1984; Yu and  
594 Hong, 2006). Longitudinal sediment transport is common in most foreland basins. A good example is  
595 the southern Pyrenees, where longitudinal sediment routing systems dominated a wedge-top depozone,  
596 with deep marine sedimentation prevailing (Mutti, 1977, Castelltort et al., 2017). It is important to  
597 note in contrast, that an averaged orogen-wide erosion rate of 3 mm/yr produces a sediment volume  
598 that is consistent with the preserved sediment volume in the western foreland basin (Fig 15, Model C).  
599 This means that, according to our approach, either previous estimates of erosion rate based on  
600 thermochronological constraints are too high, or sediment bypass occurred at least for parts of basin  
601 history.

602 Because of the presence of many submarine canyons draining sediment from the Taiwan Strait to the  
603 deeper basin in the Manila trench (Damuth, 1979; Yu and Chang, 2002; Yu et al., 2009), a  
604 fundamental unknown is whether one can find there the missing sediment volume suggested by our  
605 calculations. Sparse literature data are available on the nature of the sedimentary basins in the area of  
606 the South China Sea close to Taiwan (Lee et al., 1993; Lin et al., 2008; Yu and Huang, 2009), with a  
607 main focus on the Pearl River delta and associated submarine fan deposits (Lüdmann et al., 2001; Su  
608 et al., 1989; Xiong et al., 2004) (Li et al., 2008). A topographic map of the submarine regions south of  
609 Taiwan indicates a peculiarity in the slope of the South China Sea continental margin compared to its  
610 continuation further to the South. This suggests an anomalous accumulation of sediment in this area.  
611 Topographic profiles across and along the continental margin (Fig. 16, inset) show that the ocean floor  
612 remains at a bathymetry of about -4000 m. As a first order approximation we use the isobath -3600 m  
613 and a line roughly parallel to the shelf edge to delimit the contour of this promontory of the continental  
614 margin and to compute its volume. The volume enclosed by the area drawn on Figure 16 and using -  
615 4000 m as a base elevation represents 28'700 km<sup>3</sup> (15'400 km<sup>3</sup> when -3600 m is used as a base  
616 elevation for the calculation).





617  
 618 **Figure 16: Digital elevation model of the South China Sea with topographic profiles across the submarine promontory**  
 619 **of the -3'600m isobath.** The volume stored in the shaded area (contoured with dashed line) represents  $\sim 30'000 \text{ km}^3$ .

620

621 The volume of this submarine topography is compatible with deposits originated in the Taiwan  
 622 orogeny that would have bypassed the Taiwan Strait. The outline of Tainan basin on the topographic  
 623 map of figure 16 and its southwestward orientation visible on paleogeographic maps of figure 5 show  
 624 that Tainan basin may have constituted a longitudinal through working as a conduit for material  
 625 sourced in Taiwan orogenic wedge. In this case, a non-negligible portion of the sedimentary record of  
 626 mountain building may have been preserved outside of the foreland basin itself. However this  
 627 hypothesis remains to be tested with future work investigating the sedimentological nature and  
 628 stratigraphy of this anomalous promontory and look for potential sediment depocenters outside of the  
 629 Taiwan Strait. This finding outlines the potential complexity of interpreting provenance signals  
 630 (Romans et al., 2016) in orogen-basin systems with highly dynamic topographic evolution.

631

## 632 5. Conclusions

633 The sedimentary system of the Taiwan foreland basin is governed by oblique collision between Luzon  
 634 volcanic arc and the Asian passive margin. Different geometrical models of orogen growth and its  
 635 influence on basin architecture were tested by means of a stratigraphic modeling approach. By looking  
 636 at sediment volumes in the foreland basin and calculating mass flux sediment budgets, we document  
 637 that a significant (perhaps more than 50%) portion of the sediment eroded from the orogen is not

638 preserved in the stratigraphic record of the immediately adjacent foreland basin. The excess sediment  
 639 is most likely transported northwards into the Okinawa Trough and southwards into the South China  
 640 Sea, where large submarine channel-lobe systems developed. This interpretation is consistent with an  
 641 increasing amount of submarine incisions since Late Pliocene observed in southwest Taiwan.  
 642 We propose that this may explain why, despite being in front of a steady state orogen, the basin  
 643 remains in an "underfilled" state. We tested three different orogenic growth scenarios with  
 644 longitudinal transport. While predicted preserved sediment thicknesses exceeded observed sediment  
 645 thickness, longitudinal transport was efficient enough to keep the basin from overflowing in all three  
 646 scenarios. However, we find that, despite recent suggestions that collision in Taiwan may have been  
 647 synchronous along its entire length (over the length of Taiwan, Castelltort et al., 2011, Lee et al.,  
 648 2015), an oblique collision fits better the observed basin architecture.

649

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