Variations in Inelastic Failure of Subducting Continental Lithosphere and Tectonic Development: Australia-Banda Arc Convergence

Kush Tandon[1], Juan M. Lorenzo[2], Sri Widiyantoro[3], and Geoffrey W. O'Brien[4]

Factors that control the geometry of foreland basins and accretionary prisms, back thrusting, and changes in structural style during the continental collision can be linked to variations in Effective Elastic Thickness (EET). A variable EET map is computed at an incipient continental collision (Pliocene-Recent) site in the northern Australian continental lithosphere along the Banda orogen. Incorporation of almost all the forearc basin within the central Timor Island and tomogram in the eastern Timor Island suggest a more rigid northern Australian lithosphere indenting between ~125°-127° E longitude. Modeled flexure deflection is matched to seafloor bathymetry and Banda orogen topography, Pliocene continental shelf, and marine complete 3D Bouguer gravity anomalies. Calculated EET of the northern Australian continental lithosphere using an elastic half-beam model show a sharp decrease in EET from 230-180 km on the continental shelf (from Roti to west of Aru Island) down to ~40 km on the continental slope and beneath Banda orogen (from ~ 121°-132° E longitude) favoring an hypothesis of inelastic failure at the start of continental subduction. The decrease in EET along the continental slope is not entirely uniform and parallel to the subduction boundary, especially south of central Timor Island. In the south of central Timor Island, the EET remains close to 55 km on the continental slope and beneath the central Timor Island. An increase of ~15 km in EET calculations on the continental slope south of central Timor Island shifts the majority of present-day strain partitioning to the forearc region as compared to regions in western Timor Island and Tanimbar Island. A pre-existing stronger region of continental lithosphere present within the transfer zone between Malita and Valcun Graben (continental shelf) is a likely cause for lesser inelastic failure in that region during subduction. Results show that continental lithosphere is inherently strong on the continental shelf under normal crustal thickness and in absence of significant weakening from past rifting. Variations in the gradient of EET change on the continental slope can determine the evolution of many features in continental collision tectonics.

Plate Boundary Zones
Geodynamics Series 30

10/1029/030GD20

[1]Department of Geology and Energy Institute, University of Louisiana, Lafayette, Louisiana
[2]Department of Geology and Geophysics, Louisiana State University, Baton Rouge, Louisiana
[3]Department of Geophysics and Meteorology, Bandung Institute of Technology, Bandung, Indonesia
[4]Australian Geological Survey Organization, GPO Box 378, Canberra, Australia

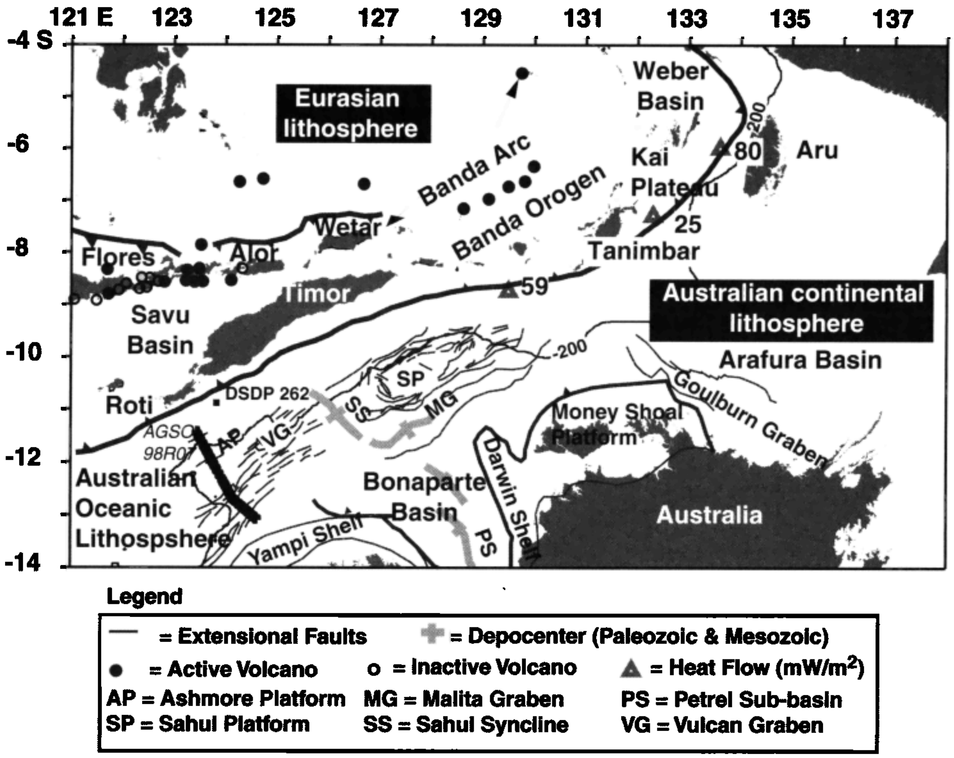

Figure 1. General tectonic map for the northern Australian continental lithosphere across the Banda Arc [Hamilton, 1974; Veevers et al. 1974; Stagg, 1993; AGSO North West Shelf Study Group, 1994; Snyder et al., 1996]. Heat flow values are from Bowin et al. [1980]. The thick line with black triangles (point to overthrust plate) shows the subduction boundary between the northern Australian continental lithosphere and the Eurasian lithosphere in Timor-Tanimbar-Aru Trough; north of Flores, in Alor and Wetar, it denotes the back-thrusting. Thick black line lettered "AGSO 98R07" marks the location of seismic track for AGSO line 98R07. The 200 m bathymetry contour marks the continental shelf-slope boundary from ETOPO-5 [National Geophysical Data Center, 1988].

1. INTRODUCTION

Local area Airy isostasy (hydrostatic equilibrium) is generally found only in large-scale geological features, such as Tibetan Plateau [Lyon-Caen and Molnar, 1985; Burov and Diament, 1992 and 1996]. In Himalayas and other mountains chains, the flanks of the mountain ranges show rigid behavior as opposed to local area Airy isostasy [Lyon-Caen and Molnar, 1983 and 1985; Royden and Karner, 1983; Stewart and Watts, 1997]. This enables a part of the continental collision margin to support the weight of the thrust sheets and lead to formation of foreland basins [Lyon-Caen and Molnar, 1983 and 1985; Royden and Karner, 1983; Stewart and Watts, 1997]. Within the same continental collisional margin, areas with similar convergence vector also display markedly different tectonic development (Figure 1). Differences in the rigid part of the continental lithosphere along a collisional margin, especially in the down dip direction should play an important role in determining the progress of the continental collision tectonics and how foreland basins evolve with time.

Effective Elastic Thickness (EET) measures the ability to support loads and transmit bending stresses regionally as a rigid lithosphere rather than mass compensation being in a local Airy isostatic manner (zero EET) [McAdoo et al., 1978; Watts, 1983]. There has been a long-standing debate on factors that control the value of EET in the continental lithosphere, like inelastic failure, crustal thickening, and tectono-thermal age (the time elapsed between the last rifting event and the start of formation of a foreland basin) [McNutt et al., 1988; Burov and Diament, 1995 and 1996; Stewart and Watts, 1997; Poudjom Djomani et al., 1999].

Watts et al. [1995] advocate that the highest the value of EET should be found where the collision has maximum shortening. Pre-collisional high EET facilitates thin-skinned, low-angle fold and thrust tectonics, and back-thrusting on a low-angle flexurally more rigid plate [Watts et al., 1995]. EET values differing by 1.5 times within a 200 kilometers range are noticed in the Himalayas and Ganga basin [Lyon-Caen and Molnar, 1985]. In the Apennine foreland basin system, Royden et al. [1987] showed that the differences within subduction processes, foreland basin and thrust sheet geometry can be related to the segmentation of the lithosphere with uniform EET. In the central Brazilian Shield, Stewart and Watts [1997] calculated variable EET of > 85 km on the craton that decreases to < 25 km beneath the thrust sheets. In the central Brazilian Shield [Stewart and Watts, 1987], the EET variations are not uniform and not strike parallel to the thrusting.

At a site of a Pliocene-Recent continental-arc collision, the northern Australian continental lithosphere along the Banda orogen [Warris, 1973] one observes significant variations within different elements of plate-boundary interactions, e.g., strain partitioning, structural style, geometry of the foreland basin and the adjoining accretionary prism, and presence of back-thrusting. In this study, we attempt to find a relationship between variations in EET (down dip and along the strike of the thrust sheets) and different plate boundary observations found during the juvenile stages of continental collision tectonics. We perform elastic half-beam modeling along 15 profiles (Figure 6) to calculate variable EET from Roti to the Kai Plateau (west of Aru Island) traversing through Timor and Tanimbar Islands (~ 121°-137° E longitude) (Figure 1). Modeled flexure from elastic half-beam modeling is matched to the seafloor, the Banda orogen, the continental shelf from Pliocene, and the marine complete 3D Bouguer gravity anomalies. Variable EET on the continental slope accounts for the change in the mechanical strength of the continental lithosphere down dip caused by inelastic yielding [McNutt et al., 1988].

2. DIFFERENT FEATURES OF CONTINENTAL COLLISION TECTONICS

At the start of continental subduction, there are normal faults due to bending stresses on the continental shelf and slope [Bradley and Kidd, 1991], thrusting within the accretionary prism [Karig et al., 1987] and extension in the forearc basin [Uyeda and Kanomori, 1979]. However, the seismic reflection data from the northern Australian continental shelf-slope (Figure 1) show that the extent of normal faulting is reduced significantly in Late Miocene-Early Pliocene strata on the continental shelf south of Timor Island (Figure 2). The Late Miocene-Early Pliocene is generally thought of as the time when continental subduction started in northern Australia [Veevers et al., 1978; Johnston and Bowin, 1981]. Regionally, the timing of termination of the majority of normal faults migrates from Late Miocene-Early Pliocene-Recent as one moves from west of Timor Island toward Tanimbar Island [Curry et al., 2000] (Figure 3). Interestingly, the strain partitioning in the forearc basin north of central and eastern Timor Island is under compression unlike north of western Timor and Tanimbar Island [Charlton, 1997] (Figure 3). Foreland vergance has stopped at present in central and eastern Timor Island but is still active in western Timor and Tanimbar Islands [Charlton, 1997] (Figure 3). The width of the forearc basin north of central Timor Island is also the least compared to other regions north of the Banda orogen (Figure 1). In the vicinity of Timor Island, the bending stresses caused by continental subduction are neutralized by horizontal compression as soon as the subduction started. On the contrary, near Tanimbar Island, the bending stresses still exceed the horizontal compression (Figure 3).

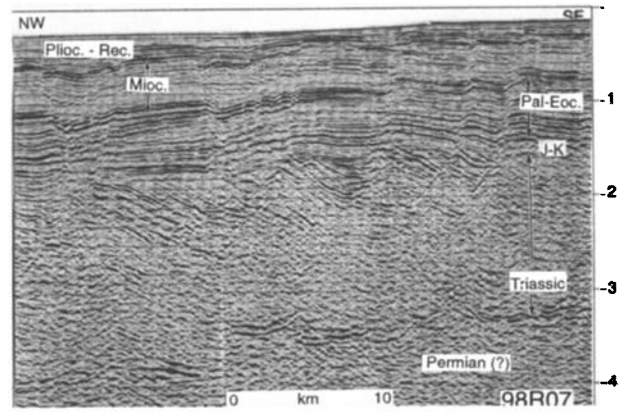

Figure 2. Uninterpreted migrated seismic reflection AGSO line 98R07 on the northern Australian continental shelf southwest of Timor Island. Age abbreviations: Plioc(ene)-Rec(ent), Mioc(cene), Pal(eocene), Eo(cene), J(urassic)-K(Cretaceous), Tr(iassic), and P(ermian). The seismic section shows how the extent of normal faulting changes with time. Some faults appear to detach within Cretaceous units. Most reactivated faults terminate above, or near, a regional base Pliocene unconformity [Boehme, 1996; Curry et al., 2000] that constrains the end of the dominant extensional stress regime to Latest Miocene/Earliest Pliocene. The location for AGSO line 98R07 is shown in Figure 1.

The P-seismic wave tomography indicates that the north-northern Australian continental lithosphere north of eastern Timor Island is more detached from the leading edge of the subducting Australian lithosphere than near western Timor and Tanimbar Island (Figures 3 and 4). In addition, central Timor Island has highest topographic elevation

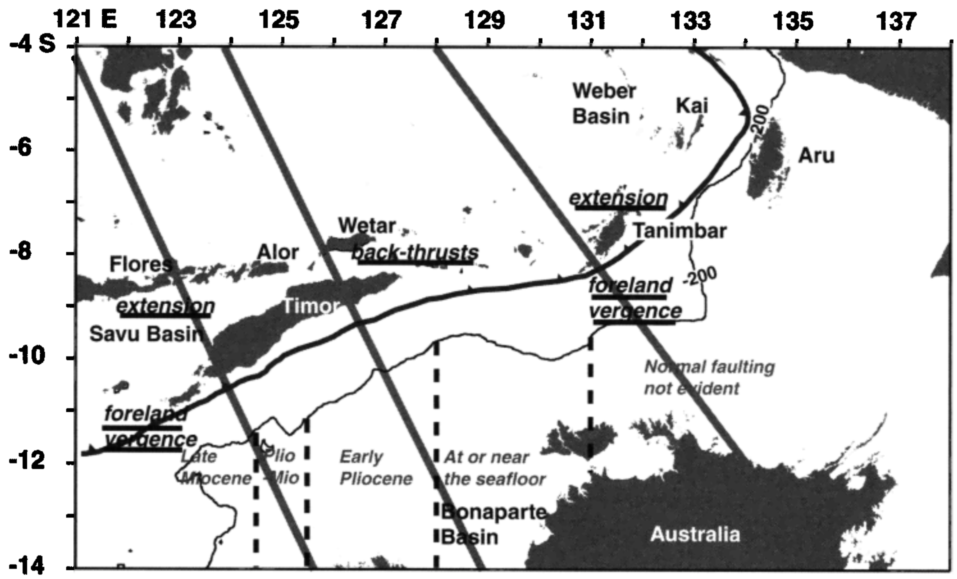

Figure 3. Present-day dominant structural style on the continental shelf, near the subduction boundary and in the forearc basin. Age of termination of majority of normal faulting on the continental shelf of the northern Australia is mapped by thousands of kilometers of seismic reflection data by Curry et al. [2000] (shown as italic on the continental shelf with vertical dashed line as boundaries). There is still evidence for present-day extensional faulting (caused by bending stresses) as seafloor scarps on the northern Australian continental slope near Timor Island but with a reduced areal extent in the shelf-slope compared to Late Miocene-Early Pliocene [Boehme, 1996]. Different regions of present-day strain partitioning near the subduction boundary and in the forearc basin are shown in underlined bold italics [Charlton, 1997]. The thick NW-SE lines (a, b, c in Figure 4) are the profiles for seismic tomograms.

compared to the other islands in the Banda orogen (Figure 5). These observations seen on the northern Australian continental shelf, Timor-Tanimbar-Aru Trough, Banda orogen and Banda Arc could be responding to regional changes in EET.

3. WORKING HYPOTHESIS

Quite possibly, the cessation of normal faulting seen at Miocene-Pliocene on seismic reflection data (Figure 2) is because the extensional stresses were greater on the northern Australian continental lithosphere when it was far away from the continental subduction compared to its position today near the shelf-slope boundary. One of the possible mechanisms for reducing extensional stresses is decreasing the bending stresses [Turcotte and Schubert, 1982]. Bending stresses can be reduced by decreasing the EET (less rigid) of continental lithosphere at the start of the continental subduction and/ or the decrease the curvature of the lithosphere (less bent) [Turcotte and Schubert, 1982].

Even the decreased bending stresses need to be neutralized by increased horizontal compression within northern Australian shelf and Timor-Tanimbar trough for normal faulting to stop (Figure 2). A more rigid northern Australian continental lithosphere on the continental slope and beneath the orogen (higher EET values) will promote thin-skinned tectonics on an indenting surface that supports more loading. That may cause increased horizontal compression, as seen in the vicinity of central Timor Island. However, higher EET values will also have more bending stresses compared to areas of lesser EET values. Another mechanism for increasing horizontal compression during collision is break-off of the lithosphere within the subducted part [Chemenda et al., 2000].

Higher EET values on the continental slope will resist subduction and can shift the locus of present-day strain accommodation to the forearc region during the continental-arc collision. Similarly, higher EET continental lithosphere beneath the orogen can also aid in the detachment of the shallow parts of continental lithosphere from the oceanic lithosphere already subducted. A buoyant, detached continental lithosphere could stop the subduction.

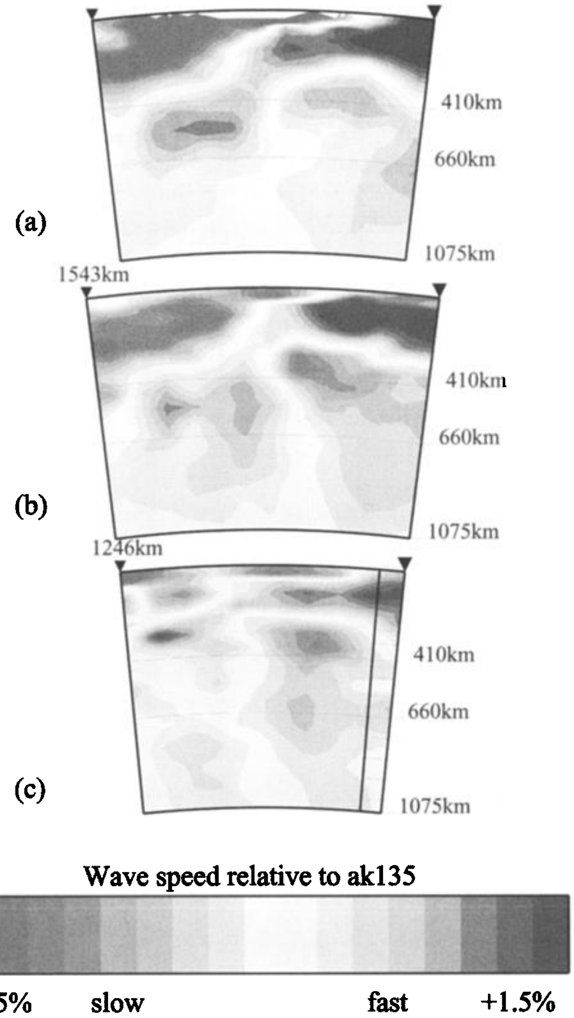

Figure 4. Vertical tomograms across: a) west of Timor Island, b) through the eastern Timor Island, and c) Tanimbar Island from a P-wave tomographic model by Widiyantoro and van der Hilst [1996]. The tomograms are plotted from the north (left) to the south (right) in terms of wave speed perturbations relative to ak135, a radially stratified velocity model [Kennett et al., 1995]. The blue color is the inferred subducting northern Australian lithosphere. Locations are shown in Figure 3.

4. GEOLOGICAL BACKGROUND

The northern Australian continental lithosphere started continental subduction across the Banda orogen in Late Miocene-Early Pliocene [Veevers et al., 1978; Johnston and Bowin, 1981]. The earliest record of collision between Australian and Eurasian lithosphere in Timor Island is marked by the presence of ~38 Ma (Late Eocene) metamorphic overprint on ophiolites [Sopaheluwaken et al., 1989]. The northern Australian continental lithosphere also underwent NW-SE striking Late Devonian-Early Carboniferous rifting (Petrel sub-basin in Figure 1) and ENE-WSW striking Jurassic (Malita graben in Figure 1) [Harris, 1991]. At DSDP 262 site in the Timor Trough (Figure 1), the transition from shallow-water to deep-water sedimentation in the Early Pliocene indicates that the subduction of northern Australian continental lithosphere started at least 3 Ma ago near western Timor Island [Veevers et al., 1978; Johnston and Bowin, 1981].

The northern Australian continental lithosphere south of Roti near Ashmore Platform (Figure 1) is now estimated to be converging with the Eurasian lithosphere [Warris, 1973] at a rate of 74 +/- 2 mm/ yr in a N 17 °E +/- 3° direction [DeMets et al., 1994] (Figure 1). Analysis of GPS [Genrich et al., 1996] and earthquake seismology [McCaffrey and Nabelek, 1986] data also show that at present the subduction within the Timor Trough has stopped and the majority of strain is being accommodated within the forearc basin and the back-thrust faults [Kreemer et al., 2000] (Figure 1). Recent GPS calculations near Darwin (Australian Shelf) indicate that the northern Australian continental lithosphere-Eurasian lithosphere as a block is converging with regard to the Sunda shelf at a 7% slower rate compared to DeMets et al.'s [1994] estimation but in the same direction [Genrich et al., 1996].

Earthquake foci north of the Banda orogen at intermediate depths (50-100 km) also indicate that the leading edge of the Australian lithosphere is presently detaching at those depths [McCaffrey et al., 1985; McCaffrey, 1988; Charlton, 1991], as also seen in tomograms (Figure 4). The isostatic rebound caused by partial decoupling of the Australian lithosphere is evidenced by uplift of the Banda orogen and the Banda Arc islands [DeSmet et al, 1990]. The tomograms (Figure 4) do not indicate complete detachment of the leading edge of the northern Australian lithosphere.

5. DATA

Modeled flexure is matched to ETOPO-5 bathymetry data that is gridded at 5 minute (~ 9 km in the area of study) intervals [National Geophysical Data Center, 1988] (Figure 1). We did not use Smith and Sandwell's [1997] bathymetry database as a very high-resolution bathymetry converted from gravity can run into nonlinear problems on the continental shelf [Mackenzie, 1997]. More importantly, we want to keep the bathymetry [National Geophysical Data Center, 1988] and gravity database [Sandwell and Smith, 1997] independent of each other. A sedimentary surface marking the base of Plioquaternary sediments [Curry et al. 2000] from Australian Geological Survey

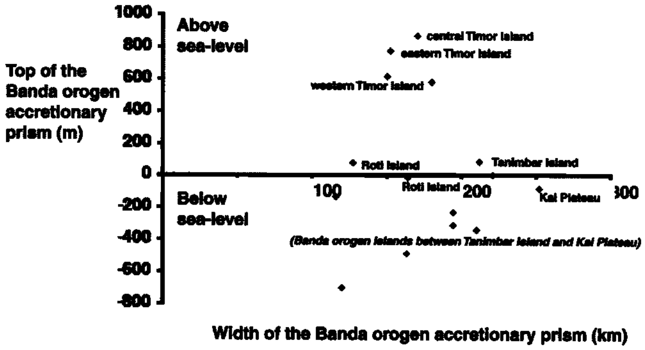

Figure 5. The width of the Banda orogen measured from the back-stop to the Timor-Tanimbar-Aru Trough and the maximum topographic elevation within the Banda orogen measured from the 15 profiles (transects shown in Figure 7) used for elastic half-beam modeling. Note that the central and eastern Timor Island has the maximum elevation compared to other areas and also shows evidence that the area (the Wetar thrust) is becoming a doubly-vergent orogen [Kreemer et al., 2000].

Organization (AGSO) seismic reflection surveys (Figures 6 and 7) is also used.

Satellite-derived marine free-air gravity anomaly data [Sandwell and Smith, 1997] across the Banda Arc are used to obtain complete 3D Bouguer gravity anomalies (Figure 8). A 3D Bouguer gravity anomaly corrects for limited length-scale of the density contrast along the strike direction [Blakely, 1995]. Whereas, 2D Bouguer anomaly assumes that the anomaly being caused by the density contrast is infinite in extent along the strike direction [Blakely, 1995]. Sandwell and Smith's [1997] data is prepared from Geosat Geodetic Mission and ERS 1 Geodetic Phase with a horizontal resolution of ~10 km. The accuracy of the worldwide satellite-derived free-air gravity anomaly is 3-7 mgal [Sandwell and Smith, 1997]. Marine complete 3D Bouguer gravity anomalies are created from the satellite-derived free-air gravity anomalies [Sandwell and Smith, 1997] by replacing the seawater with Tertiary sediments (Table 1). The 3D bathymetric corrections [Blakely, 1995; Tandon, 2000] are applied using ETOPO-5 [National Geophysical Data Center, 1988].

A seismic tomogram based on the regional model for Indonesia developed by Widiyantoro and van der Hilst

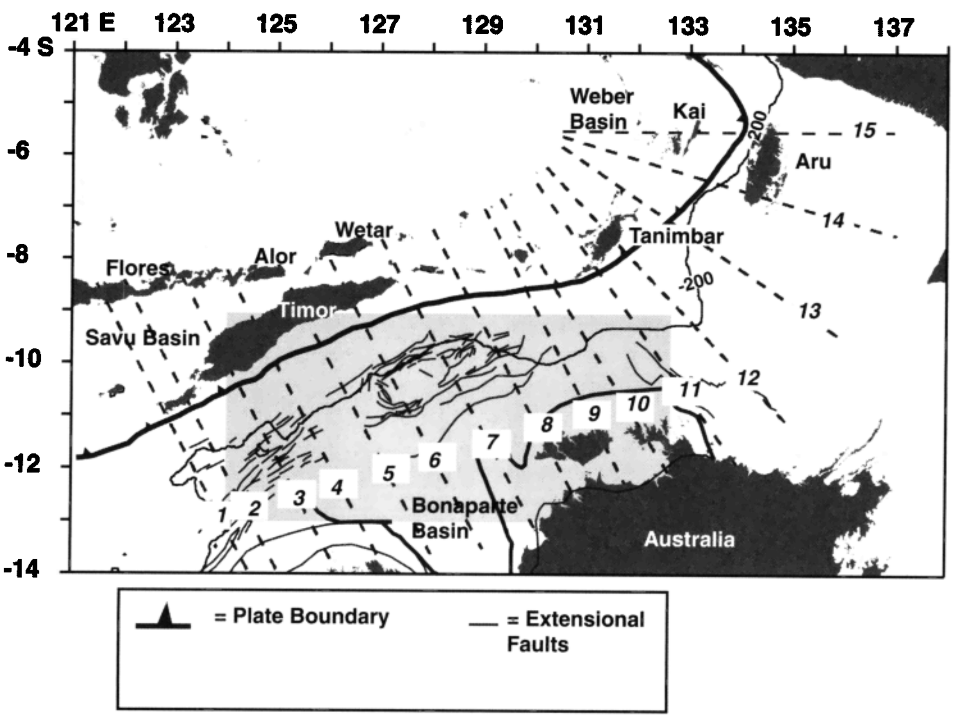

Figure 6. The dashed lines locate numbered transects used for the elastic half-beam modeling for variable Effective Elastic Thickness (EET) modeling. The box shows the region where we reconstruct the continental shelf during Pliocene to be used for flexure modeling in addition to seafloor bathymetry, Banda orogen topography and marine complete 3D Bouguer gravity anomalies.

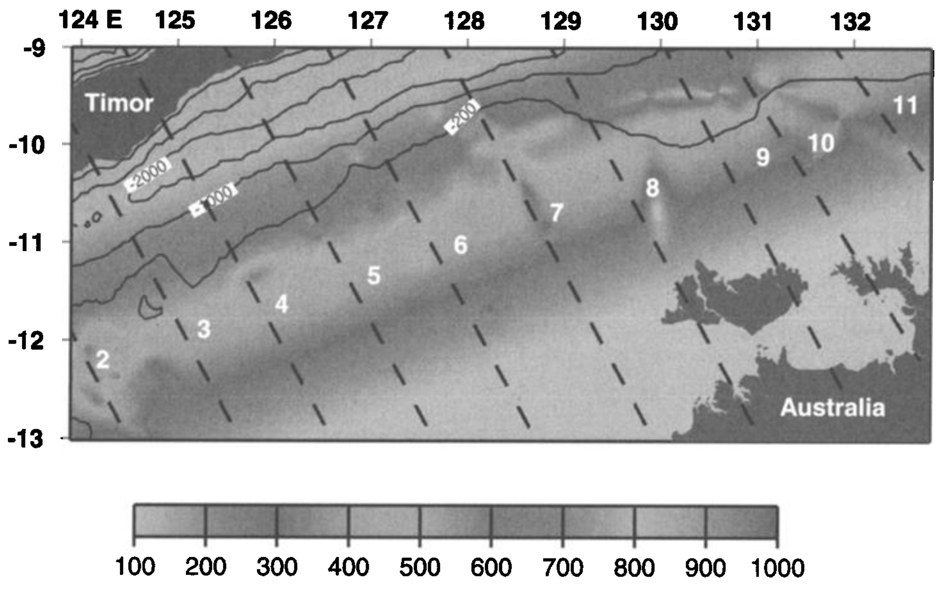

Figure 7. Gridded map of the Plioquaternary sedimentary surface in mbsf (meters below sea floor) created by mapping thousands of kilometers of seismic reflection data [Curry et al., 2000]. Numbered dashed lines are transects for elastic half-beam modeling within that region. The contours are for seafloor bathymetry. The thickness of Plioquaternary sediments is determined by subtracting the base Plioquaternary (sedimentary) surface from the seafloor bathymetry.

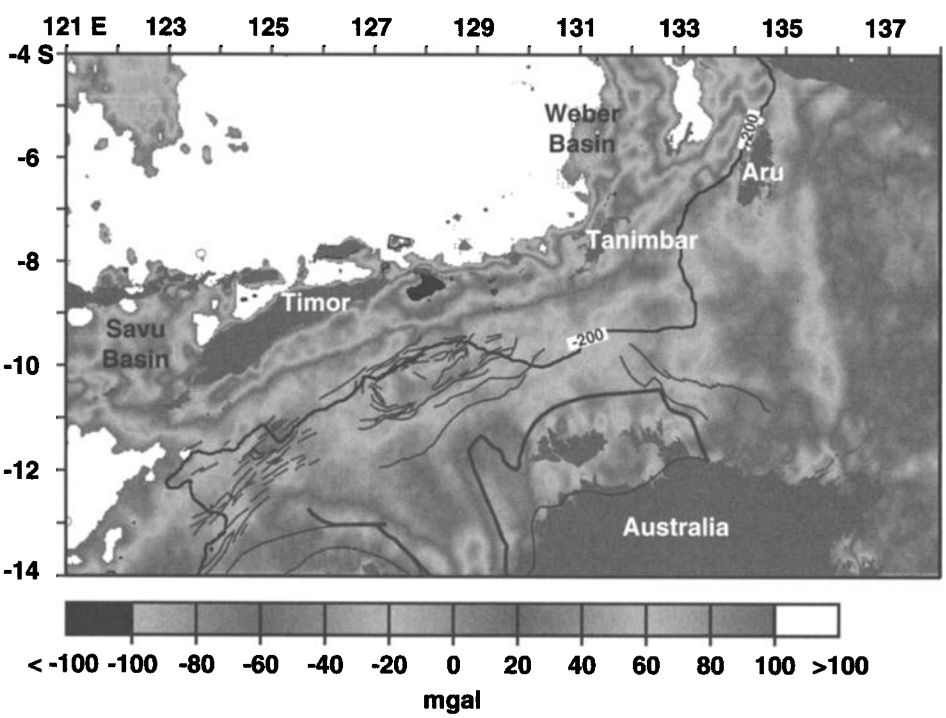

Figure 8. Marine complete 3D Bouguer gravity anomalies for the northern Australian continental lithosphere across the Banda orogen. The gray shaded area corresponds to the portion of the Banda Arc above the sea level with no marine satellite data coverage [Sandwell and Smith, 1997]. Geological information overlain is from Figure 1.

Table 1
Parameters for Flexure Modeling

Name	Value
Parameters for Elastic Half-Beam Modeling	
Elastic Beam	
Young's Modulus (E)	5 X 10^{10} kg/ ms^2
Poisson's Ratio (γ)	0.25
Timor-Tanimbar-Aru Trough (Sea water)	
Density	1030 kg/ m^3
Supporting Fluid (Mantle)	
Density	3300 kg/ m^3
Triangular Wedge (Banda orogen)	
Density	2670 kg/ m^3
Parameters for Gravity Modeling	
Density	
Sea water	1030 kg/ m^3
Tertiary sediments	2200 kg/ m^3
Banda orogen	2670 kg/ m^3
Australian continental crust	2800 kg/ m^3
Mantle	3300 kg/ m^3

Table 1. Parameters [Chamalaun and Grady, 1976; Turcotte and Schubert, 1982; Richardson, 2001] used for a) elastic half-beam, and b) gravity modeling. All the units are defined in SI.

[1996] is shown in Figure 4. The model has been produced by a cellular representation of the mantle structure. The mantle volume inside and outside the study region was parameterized using cells of 1.0^{o} x 1.0^{o} and 5.0^{o} x 5.0^{o}, respectively. The above mentioned model parameterization allows the resolution of relatively small-scale structure in the region under investigation and minimizes contamination from outside. The P-wave delay time data set inverted to construct the tomographic model is from the global data processed by Engdahl et al. [1998].

6. ASSUMPTIONS

6.1 Elastic Half Beam Models

We approximate the northern Australian continental lithosphere with a 2D elastic half-beam model of variable effective elastic thickness (EET) embedded in an inviscid fluid [Turcotte and Schubert, 1982] (Figure 9). A laterally variable EET model incorporates the decrease in EET due to inelastic processes, such as brittle and plastic failure, and mechanical decoupling between upper-lower crust and mantle [McNutt et al., 1988].

High EET values on the continental shelf combined with EET values decreasing down dip on the slope will have a forebulge with lower amplitude compared to a constant high EET estimate. As it is, a constant high EET estimate (> 80 km) creates a shallow but long wavelength forebulge on the continental shelf that can be further decreased if the value of EET decreases down dip. Decreasing EET on the continental slope due to inelastic failure decreases the capacity of the continental lithosphere to transmit stresses and does not produce idealized forebulges [McAdoo et al., 1978]. In contrast, a constant low EET (~20-40 km) value creates a low wavelength, high amplitude forebulge on the shelf. Our observation shows lack of high amplitude forebulge seen on the seafloor bathymetry [National Geophysical Data Center, 1988] and from Pliocene continental shelf on the continental shelf of northern Australia [Lorenzo et al., 1998; Tandon et al., 2000]. Evidence for significant reduction from high EET values on shelf to low on the slope is the presence of high curvature (~10^{-7} m^{-1}) on the slope along with no clear evidence for a small

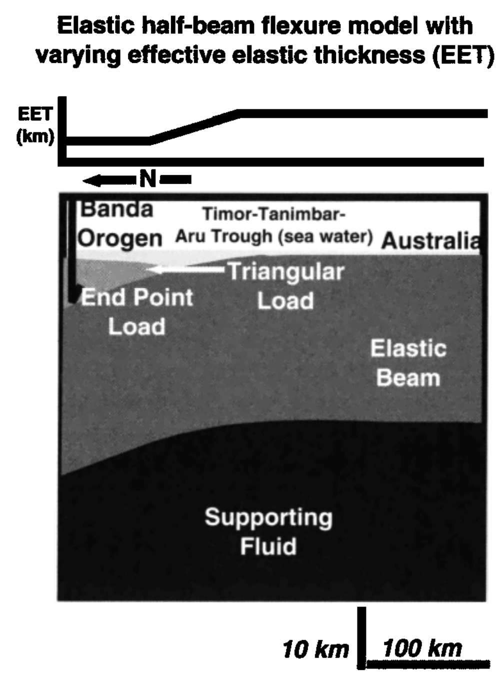

Figure 9. An elastic half-beam model with laterally variable Effective Elastic Thickness (EET) is used to calculate the EET of the northern Australian continental lithosphere. The Banda orogen in the elastc half-beam model is approximated by a triangular load. The surfacial deflection created by the elastic half-beam model is filled with sea water. The end-point load at the northern end of elastic half-beam approximates the subsurface component of loading. Laterally variable EET model has the EET decreasing from Australia to Timor-Tanimbar-Aru trough to Banda orogen assuming inelastic failure.

wavelength forebulge on the shelf [Lorenzo et al., 1998]. Lack of significant forebulge is not due to horizontal tensile forces [Bowin et al., 2000] because the stress regime on continental shelf does not show extension being dominant today (Figures 2 and 3).

The northern Australian continental lithosphere since Pliocene is assumed to bend under the weight of the Banda orogen acting as a surface load (Figure 9). Ideally, using the Plioquaternary sediments from shelf to slope along with Banda orogen sediments would be a more accurate description of surface load. However, Plioquaternary sediments are only few hundred meters thick [Veevers, 1974] compared to 10s of kilometer thick Banda orogen [Snyder et al., 1996]. The Banda orogen acts as a major contributor to the surface loading. We only have limited information regarding the thickness of Plioquaternary sediments on the shelf when compared to the total area covered by our study (Figure 6). We do not have any detailed sedimentary thickness information on the slope. To perform a regional EET analysis from Roti to Kai plateau, we need to be consistent for comparison. In elastic half-beam modeling, a subsurface loading component is only invoked when surface loads alone cannot account for the deflection observed of reference horizon data. These subsurface loads originate in deeper parts of the lithosphere, as found in some other foreland basins, e.g., Apennine foreland [Royden and Karner, 1984]. A slab-pull can act as hidden subsurface load upon the northern Australian continental lithosphere. On the other hand, a complete detachment of the northern Australian continental lithosphere from the oceanic lithosphere would mean that there is no slab pull. A partial detachment or buoyancy of the continental lithosphere would cause the effect of slab pull to be buffered. A buffered slab pull means that the full downward pull by the dense oceanic lithosphere is not felt by the northern Australian continental lithosphere. We invoke the concept of a buffered slab because the northern Australian continental lithosphere is assumed to be detaching from the oceanic lithosphere [Chamalaun and Grady, 1978; McCaffrey et al., 1985; McCaffrey, 1988; Charlton, 1991]. Other subsurface loads include ophiolites, like the ones found in Timor Island [Sopaheluwaken et al., 1989].

Seafloor bathymetry data allows us to match elastic half-beam modeling on a regional basis for northern Australian continental lithosphere. In an under-filled foreland basin where few hundred meters sedimentation has occurred since Pliocene, one can use seafloor bathymetry as the bent horizon for flexure modeling. In the regions where we have the information about Plioquaternary sedimentation, it is used in the modeling.

6.2 Reconstructing Continental Shelf in Pliocene

The thickness of the Plioquaternary sediments and the flexure caused by the weight of Plioquaternary sedimentation is removed from the seafloor to reconstruct the continental shelf that existed in Pliocene at the start off continental subduction. Based on present-day convergence rates between Australia and Eurasia (Figure 1), in Early Pliocene the current shelf/ slope transition was ~350 km further south and therefore was not significantly affected by the bending of the oceanic part of the northern Australian lithosphere. The thickness of the Plioquaternary sediments is calculated by subtracting the base Pliocene surface constructed from seismic surveys (Figure 7) with the seafloor bathymetry. Since, we do not have regional information about the porosity-depth relationship, therefore we do not decompact the older sediments (Miocene onwards) as done in other studies [van der Meulen et al., 2000]. We use EET = 170 km as the EET of the Australian continental shelf prior to the continental subduction which we load with Plioquaternary sediments. The choice of pre-continental subduction EET is based present-day Australian continental shelf EET values that range from 230-125 km traversing from Roti to Kai Plateau (see the Results section). The thickness of the Plioquaternary sediments and the flexure caused by it is subtracted from the seafloor to construct a Pliocene continental shelf.

6.3 Marine Complete 3D Bouguer Gravity Anomalies

Marine complete 3D Bouguer gravity anomalies (Figure 8) remove the gravity effect of seafloor-water interface [Blakely, 1995] and allows us to delineate intra-crustal heterogeneities and the shape of the Moho. The subduction of the flexed Moho at the outer trench slope and beneath the accretionary prism creates a negative Bouguer gravity anomaly on the order of 100 mgals spread over several hundred kilometers. Beneath the orogen, a high-density subsurface load can create a positive Bouguer gravity anomaly, thereby making a positive/ negative Bouguer gravity anomaly couple from the orogen to the craton [Karner and Watts, 1983].

Small-wavelength gravity anomalies over the shelf of the northern Australian continental lithosphere arise from the presence of horst and graben structures (Figure 1) and igneous rocks related to Late Devonian-Early Carboniferous and Jurassic rifting [Anfiloff, 1988]. Also, crustal heterogeneities at longer wavelengths and tectonic deformation other than continental subduction mask the gravity signature of the flexed Moho. The seismic refraction data

[Bowin et al., 1980] show that the crustal structure below the northern Australian continental shelf and slope is homogeneous. We assume that the changes in dip angle in the continental slope seen on the marine 3D complete Bouguer gravity anomalies are mostly due to variations in EET due to a simple crustal model of a flexed Moho. For intra-crustal density contrast to be incorporated in our study, one has to have a detailed structural and stratigraphic knowledge of the entire study area.

7. MODELING

7.1 Modeling Seafloor and Continental Shelf from Pliocene

We attempt to match the seafloor and the unloaded base Pliocene surface to elastic half-beam modeling (Figures 9 and 10) as follows:

1) Match the predicted flexure to the northern Australian continental slope using 15 transects (Figure 6) and to the continental shelf from Pliocene. However, we avoid any statistical criteria for our model fit, such as RMS fit or linear correlation because these error measures are dominated by small-wavelength variations in the seafloor bathymetry, Banda orogen topography and Buoguer gravity anomalies. We are attempting to match the shelf-slope-orogen on hundreds of kilometers scale to the modeled response. The choice of our variables EET values is qualitative and experiments on the uncertainties in the EET serve as error bounds.

2) Minimize the modeled forebulge. Lack of evidence for a small wavelength, high amplitude forebulge on the shelf in the data provides an important observation to match.

3) Match the shape of the triangular load on the flexed beam with the shape of the accretionary prism of the Banda orogen. The flexed profile from the elastic half-beam model is added to the triangular load to match the Banda orogen topography.

4) Determine the northern limit of the non-zero EET elastic half-beam by the total width of the triangular load that the elastic half-beam model can support. In our modeling effort, we initially attempt to match the width of the triangular load to the width of the Banda orogen loading a non-zero EET beam. The back-stop is assumed to be the northern boundary for the accrectionary prism and demarcation from the forearc basin north of Banda orogen (Figure 1). That is how we estimate the width of the Banda orogen and the starting estimate for the northern limit for our elastic half-beam model. In some cases the width of the triangular load used on the elastic half-beam model is significantly less than the observed width of the Banda orogen, like parts of Timor Island and Kai Plateau.

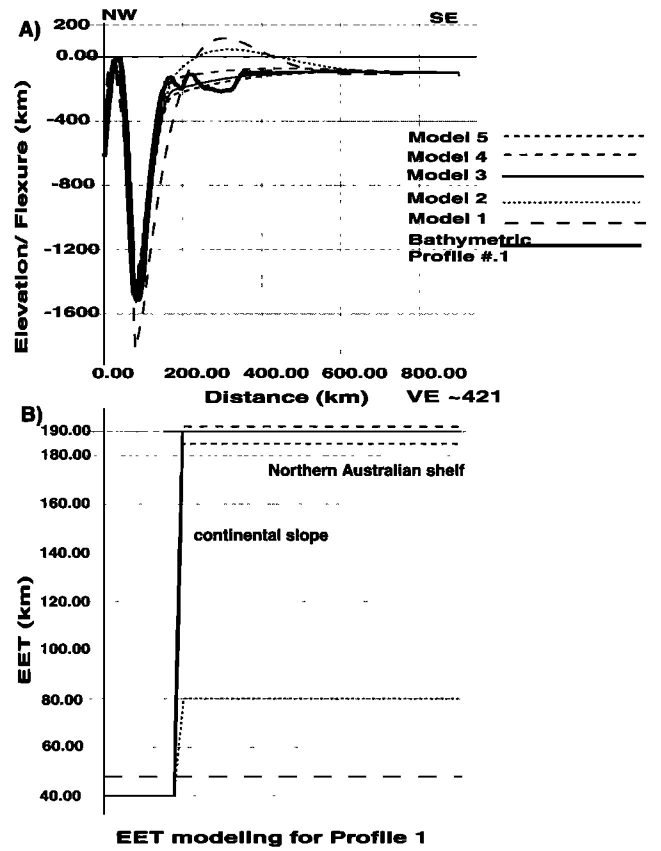

Figure 10. Increase in EET values by a factor of 2 on the continental shelf of the northern Australian continental lithosphere further decreases the amplitude of the forebulge relative to previous calculations [Lorenzo et al., 1998; Tandon et al., 2000]. Model 1 assumes constant EET down dip, Model 2 uses an EET profile similar to the ones presented by Tandon et al. [2000]. Models 3-5 are results from sensitivity analysis for the regional EET calculations presented in Figure 11. A) The top portion is the elastic half-beam response compared to seafloor bathymetry and Banda orogen topography and B) the bottom part shows EET variations for different models.

5) We start our modeling with applying a triangular load and only use an end-point load (Figure 9) at the northern end for elastic half-beam modeling if the triangular load is unable match the data. The magnitude of the end-point load is determined by the closest visual fit of the model.

7.2 Uncertainty and Sensitivity Analyses

An order of magnitude decrease in Young's Modulus (E) of the elastic half-beam results will increase the EET

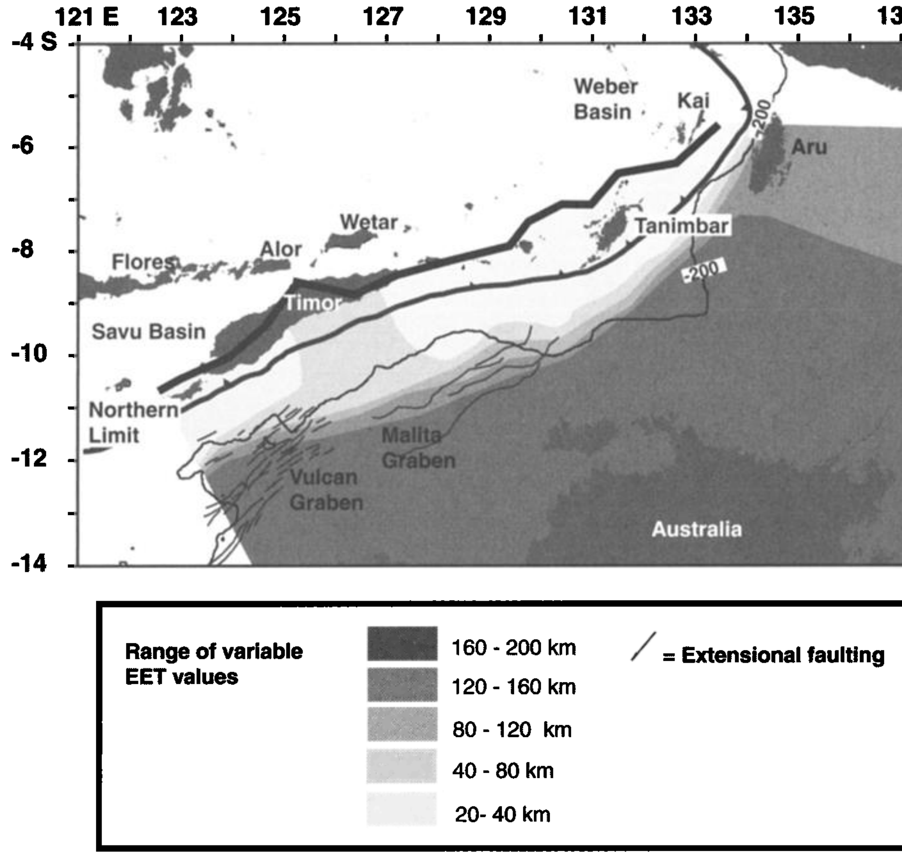

Figure 11. Contour map for variable EET calculations for the northern Australian continental lithosphere across the Banda orogen. EET values are determined by bending of the elastic lithosphere via triangular loading and end-point loading (Figure 9). The thick solid line lettered "northern limit" is the edge of the non-zero EET half-beam as determined by our model (Figure 9). The details of the loading forces are given in Figure 12. The contour interval is every 20 km. The uncertainty in EET calculations based on our sensitivity analysis is 50 km and 5 km on the slope and beneath the orogen.

by two-fold for the same flexural rigidity. However, we keep most of the parameters fixed in our modeling (Table 1). We assume that the elastic properties of the northern Australian continental lithosphere were nearly uniform along the Banda orogen prior to continental subduction. If there were significant changes in the properties of the northern Australian continental lithosphere, one would notice those changes in the morphology of the today's continental shelf-slope (Figure 1) or at the base of the Plioquaternary sediments (Figure 7).

In one of the experiments for sensitivity analysis, we varied the EET values but not the triangular load and the point load. On the Australian continental shelf, even after adding +40 km to −10 km to the EET shelf values presented in Figure 11, there is still a reasonable match to the seafloor. Interestingly, a variation of even ~5 km from EET on the Australian continental slope and beneath the Banda orogen (Figure 11) changes the modeled response appreciably compared to the data. In the modeled response, the triangular load sinks beneath the observed Banda orogen if the slope values of EET are decreased beyond 5 km. The support of Banda orogen is very sensitive to EET values on the slope and below the orogen. Our error bars have 50 km of uncertainty in EET calculations on the continental shelf but only 5 km on the slope and beneath the accretionary prism.

The sensitivity analysis for point load used in elastic half-beam modeling shows that a variation within a range of $+ 0.05 \times 10^{12}$ N/ m^2 to $- 0.05 \times 10^{12}$ N/ m^2 can still produces an acceptable fit to the data. A zero point load in Profile 5 (Figure 12) is believed to be due to lack of slab pull rather than modeling artifact. In other profiles, the

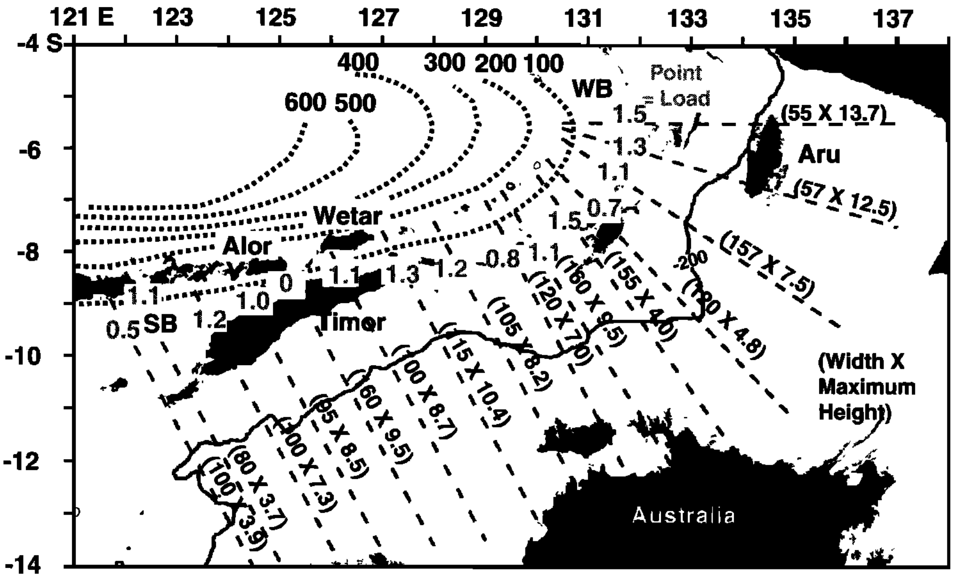

Figure 12. Forces (surface and subsurface) loading the northern Australian continental lithosphere leading to the formation of a foreland basin within the Timor-Tanimbar-Aru trough. These forces are used to compute the variable EET (km) in Figure 11. Shown within parenthesis on the northern Australian continental shelf are: a) length (km) and b) maximum width (km) for the triangular load that approximates the Banda orogen. End-point loads are used for subsurface loading are to be multiplied by 10^{12} N/m^2 (shown near the Banda orogen). SB = Savu Basin and WB = Weber Basin. The thick dashed lines are the 100 km contours for the Benioff seismic zone determined by earthquake locations [Hamilton, 1974]. Note that the point load does not correlate to the dip of the Benioff zone implying direct correlation to slab pull.

point load needed for modeled response to match the data is numerically larger than the uncertainty associated with the point load estimation.

On the eastern margin of the Banda orogen, the orocline can cause three-dimensional loading of the northern Australian continental lithosphere (Figure 1). A 2D formulation of flexure may result in an over-estimation of EET by 30 % [Wessel, 1996]. The width of the triangular load used for elastic half-beam modeling near the eastern margin of the Banda orogen is less than the width of the triangular loads used for modeling in other regions (Figure 12). Therefore, the contribution of the Banda orogen near Aru Island as a surface load is perhaps not as a cylindrical mass but more like a thin shell validating 2D approximation.

7.3 Modeling Marine Complete 3D Bouguer Gravity Anomalies

The theoretical Bouguer gravity anomalies for a flexed Moho are created from the density contrast given in Table 1. The flexure topography is added to constant thickness Moho creating crustal thickening and thinning (flexed Moho = bending of the crust by flexure + constant Moho thickness as a background). A depression in the flexure profile from elastic half-beam modeling produces an area where the crust replaces the mantle on the slope and beneath the orogen and a forebulge creates a correspondingly slight crustal thinning on the shelf. Seismic refraction data in the northern Australian continental crust along the Banda orogen [Bowin et al., 1980] provides us an estimate on the flexed Moho for gravity modeling. The crustal thickness in our model is ~31 km on the continental shelf, a lower limit from the refraction experiment [Bowin et al., 1980]. Our motivation is to investigate whether the Moho in the northern Australian shelf-slope is flexed similar to the seafloor. Therefore, the mass compensation is not in local Airy isostasy but responds to the rigidity of the continental lithosphere.

Assuming that the Moho is also flexed, theoretical Bouguer gravity anomalies are derived from the EET profiles (Figure 11) using an assumption of homogeneous crustal in outer NW Australia. Calculated Bouguer gravity anomalies are then matched to corresponding profiles from the marine complete 3D Bouguer gravity anomalies. A good fit means that the Bouguer gravity anomalies calculated from the elastic half-beam model matches the observed marine complete Bouguer gravity anomalies at the slope of the flexed continental lithosphere.

8. RESULTS

8.1 Elastic Half-Beam Deflection and Data

We show that the amplitude of the forebulge decreases further in our modeling if the EET of the northern Australian continental shelf is increased from ~80 km [Lorenzo et al., 1998; Tandon et al., 2000] to 190 km (Figure 10). Our basis for using such a high EET on northern Australian continental shelf is to minimize the modeled forebulge to match the data as much we can in the elastic half-beam modeling. Such high EET value on the shelf is plausible for normal crustal rheology [Simons et al., 2000]. Our calculations obey the thin-beam approximation (one of the assumptions for elastic half-beam modeling) since the calculated EET value on the shelf is numerically less than the wavelength of the entire flexed elastic half-beam (Figure 10) [van der Meulen et al., 2000].

The variable EET calculations by matching seafloor bathymetry generally decreases from EET = 230-180 km on the shelf to less than EET = ~ 40 km on the continental slope from Roti to west of Aru Island (Figure 11). Near Kai Plateau, the EET decreases from 155-125 km on the shelf to ~20 km on the slope (Figure 11). South of central Timor Island, the decrease in EET in northern Australian lithosphere is from 190 km to 55 km (Figures 11 and 13). Profile 5 (Figure 13c) and Profile 14 (Figure 14c) show examples of shallower and steeper failure slope compared to Profile 1 (Figure 10a). The values of EET on the slope and beneath the Banda orogen for Profiles 1 (Figure 10b), 5 (Figure 13c), and 14 (Figure 14c) are beyond the uncertainty of our modeling using our sensitivity analysis.

Regional base Plioquaternary surface (Figure 7) determined by the seismic reflection data does not show any systematic changes (migrating sedimentary depocenters) on the continental shelf-slope which may indicate evidence for deeper lithospheric processes, such as foreland segmentation [Royden et al., 1987] and/ or slab detachment [van Meulen et al., 2000]. The Pliocene continental shelf was not completely sub-horizontal as it might be a part of a long wavelength, low amplitude forebulge present during the Miocene-Pliocene boundary. That would indicate high EET values at the start of continental subduction. A small wavelength, high amplitude paleo-forebulge if formed during the Pliocene on the continental shelf has either today moved to the continental slope or been subducted beneath the Banda orogen.

Matching of the Bouguer gravity anomalies from a flexed Moho to the observed complete 3D marine Bouguer

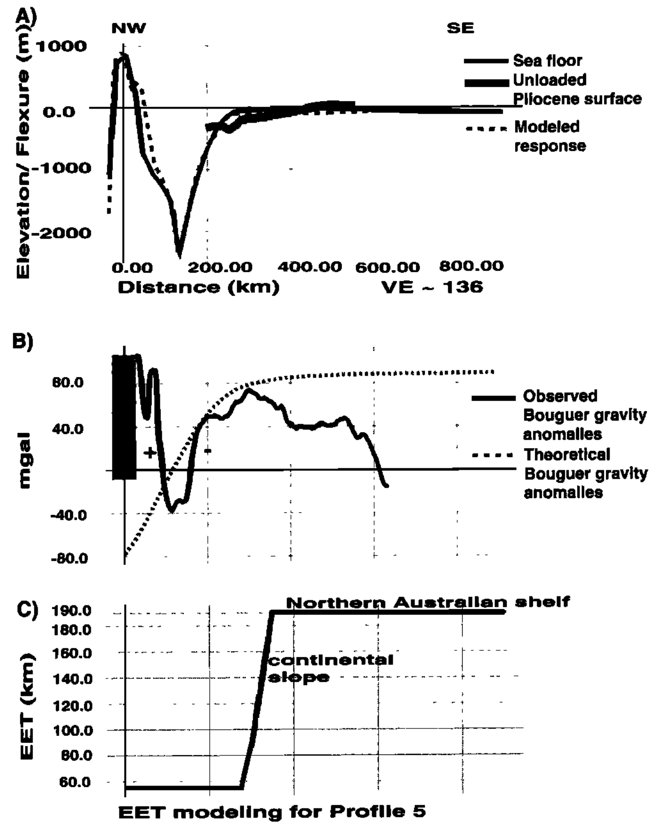

Figure 13. Variable EET values from profile 5 (Figure 6) are determined by matching the flexural profile qualitatively to the A) seafloor bathymetry and Banda orogen topography [National Geophysical Data Center, 1988) and continental shelf from Pliocene. The triangular load is added to the flexed response to match the topography of the Banda orogen. B) Theoretical Bouguer gravity anomaly created by a flexed Moho is then matched to the observed marine complete 3D Bouguer gravity anomalies. The shaded area shows that there is no observed marine Bouguer gravity anomaly data. The origin of the profile is the northern limit of the non-zero EET elastic beam for that transect (Figure 11). C) Variable EET profile. VE = Vertical Exaggeration. +/- = positive-negative Bouguer anomaly couple found traversing from the Banda orogen to the northern Australian continental slope indicates the presence of a high-density subsurface load beneath the orogen adjacent to the subducting continental crust.

gravity anomalies clearly show that Australian continental lithosphere is not in local Airy isostasy (Figures 10, 13 and 14). The heterogeneity at different scales with the crust does not allow Bouguer gravity anomalies to be used quantitatively to determine variations in EET. Matching 3D Bouguer gravity anomalies serves only as a qualitative tool testing our EET estimates using surfacial data. Figure 13b shows the limitation of simple theoretical Bouguer gravity

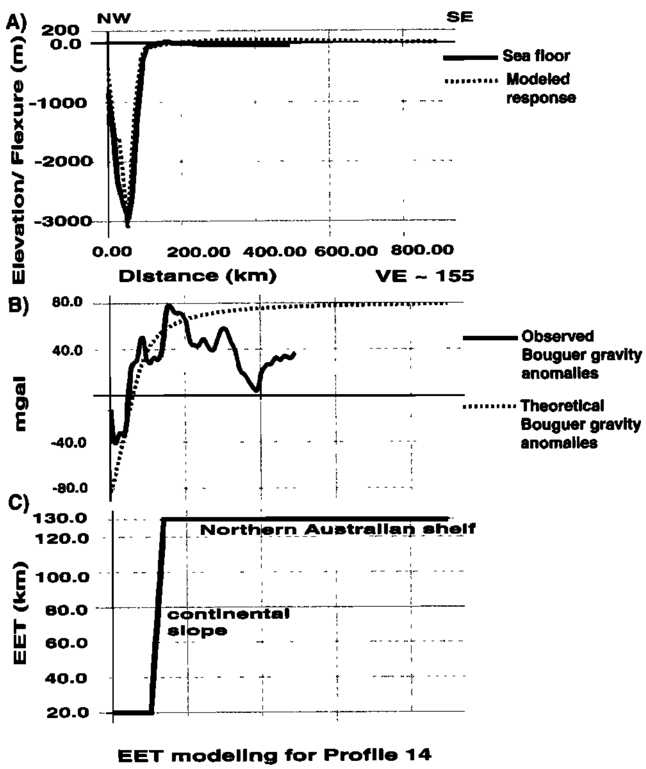

Figure 14. Variable EET values from profile 14 (Figure 6) with same description as Figure 13.

calculations. On the other hand, Figure 14b illustrates a good match between theoretical and observed Bouguer gravity anomalies even using a simple crustal model. Also, presence of positive/ negative Bouguer anomaly couple due to ophiolites in central Timor Island [Sopaheluwaken et al., 1989] is evident in Figure 13b indicates that buffered slab pull is not the only subsurface loading component in our study area

9. DISCUSSION

The changes in EET on the continental slope of the northern Australian continental lithosphere are not uniform and parallel to the subduction boundary (Figure 11). Our hypothesis is that the Jurassic rifting with ~ 20 % strain [AGSO North West Shelf Study Group, 1994] only predisposed the northern Australian continental lithosphere to weakening rather than actual weaken the lithosphere [Stewart and Watts, 1997]. Lesser decrease in EET of the northern Australian continental lithosphere on the continental slope (south of central Timor Island) is due to lithosphere's resistance to inelastic failure to some degree during the continental subduction. The transfer zone between Malita and Vulcan Graben (Figure 11) correspond to the relatively higher EET corridor from continental shelf to the slope. The Malita and Vulcan Graben were formed due to Jurassic rifting [Stagg, 1993; AGSO North West Shelf Study Group, 1994] (Figure 1). However, we do not notice regions of lesser inelastic failure at other transfer zones within northern Australian shelf. In absence of detailed publicly available structural geology map, small wavelength Bouguer gravity anomalies can help estimate positions of rift structures and transfer zones (Figure 8). Near Aru Island, the decrease in EET to 20 km (Figure 14) can be explained by an increased inelastic failure due to the oroclinal bending [McNutt et al., 1988] of the northern Australian continental lithosphere.

We notice reductions in EET close to 4-5 times in our results whereas modeling of the decoupling within the lithosphere only explains reductions of EET up to a factor of 2 [Brown and Phillips, 2000]. The bathymetric profiles (Figures 10, 13, and 14) have similar curvature seen on Tonga and Kermadac trenches [Garcia-Castellanos et al., 2000] favoring the hypothesis of a multilayered elastic-plastic rheology lithosphere that underwent inelastic failure. The results presented here are in contrast to the conclusions that the continental lithosphere have inherently low EET values of ~20 km [McKenzie and Fairhead, 1997].

Decrease in EET values from shelf to slope (Figure 11) along the entire collisional margin will decrease the bending stresses but does not explain why normal faulting has ceased near Timor but not near Tanimbar Island (Figure 3). Supporting surface loads more buoyantly near central Timor Island due to higher EET values on the slope and beneath the orogen could increase horizontal compression. Numerical models for slab break-off produce a time lag of 1.7-3.9 Ma between the start of continental subduction and the actual break-off [Wong A Ton and Wortel, 1997]. There is no evidence for complete slab break-off of the Australian lithosphere on a regional basis that would have resulted in a cessation of arc volcanism along the Banda Arc (Figure 1). A partial slab break-off on a regional scale is possible based on the evidence from seismic tomographic images (Figure 4).

If the subducting continental lithosphere has lesser inelastic failure and retains higher EET values, it not only resists subduction but perhaps also aid in the detachment of the shallower parts of continental lithosphere from the oceanic lithosphere after the continental subduction (Figures 4). Continental necking noticed on tomograms south of eastern Timor Island indicate higher possibility of detachment occurring in that area (Figure 4).

According to the calculations done by Molnar and Gray [1979], a point load of the order of 10^{11}-10^{12} N/ m^2 (Figure

13) shows that the slab pull's contribution to the formation of the Timor-Tanimbar-Aru trough (foreland basin system) is insignificant. Establishing a link between end-point load in the elastic half-beam model and deeper subduction processes using seismic tomography is still speculative. Slab pull has been identified as one of the major mechanism driving plate tectonics [Jurdy and Stefanick, 1991; Garcia-Castellanos et al., 2000]. We do not see a clear relationship between the slab pull and its contribution to the loading of the foreland basin (there is no correlation between the point load and the dip of the Benioff zone in Figure 12).

Thrusting within the prowedge and the retrowedge is noticed at all stages of the collision in clay models [Storti et al., 2000] as well as in ourt study area (Flores, Alor and Wetar back-thrust shown in Figure 1). Regional variations in EET may play an important role in causing the back-thrusting to be a dominant means of strain partitioning during the progress of continental collision (Figure 1). The height of the accretionary prism is one of the factors that controls the formation of doubly vergent orogens [Storti et al., 2000]. Higher EET values beneath the central Timor Island increases the elevation of the island (Fugure 5) and that helps in the development of back-thrusting. However, the role of EET is not clear in western part of Timor Island (Figure 11) where back-thrusting is also dominant within the forearc and the volcanic arc (Figure 1) [Charlton, 1997].

10. CONCLUSIONS

The EET changes in the down dip direction play a significant role in the evolution of continental collision tectonics. Our results show a similar relationship between the variable EET values and shortening, as also conceptualized in the central Brazilian Shield [Watts et al., 1995; Stewart and Watts, 1997]. Inherent structure of the lithosphere prior to subduction, like the presence of transfer zones from previous rifting can affect the degree of inelastic failure. The Australian continental lithosphere prior to continental subduction had an EET of at least 230-180 km (Roti Island-west of Aru Island) that decreased 4-5 times due to inelastic failure at that start of continental subduction. Continental lithosphere under normal crustal thickness and not weakened significantly by prior rifting can be very rigid. Almost the lack of forearc basin north of central Timor Island is due to increased shortening by higher EET lithosphere on the slope and beneath the orogen. Higher EET on the slope and beneath the orogen is capable of supporting more surface load and facilitating the formation of doubly-vergent orogens. On the continental slope and beneath the orogen, even an EET increase of ~15 km shifts the present-day strain partitioning to the forearc basin. Comparing end-point load in elastic half-beam modeling and seismic tomography leads us to speculate that slab pull does not contribute to the formation of Timor Trough in the vicinity of central Timor Island.

Acknowledgments. EET modeling was kindly given to us by Jonathan Stewart (code originally written by Bodine). Some of the figures are made using the GMT software [Wessel and Smith, 1995]. Comments from anonymous reviewers and the editor Seth Stein greatly improved this manuscript. Thoughtful readings of the drafts by Uwe Strecker were very beneficial. We acknowledge Jamie Curry for providing us the interpreted top of Plioquaternary sedimentary surface for our study. AOA Geophysics Inc. and LR software helped KT with gifting their software for gravity modeling. Tommy Smith from Louisiana State University helped immensely in letting use their workstation and computer programs remotely required for the research presented in this manuscript. KT acknowledges the facilities used for the writing of this manuscript and computer support by Kyle Hubbard during his stay at the Laboratory for Computational Geodynamics, Department of Chemistry, Indiana University.

REFERENCES

AGSO North West Shelf Study Group, Deep reflections on the north west shelf: changing perceptions of basin formation, in: *The North West Shelf, Australia: proceedings of petroleum exploration society of Australia symposium*, edited by P. G., and R. R. Purcell, 63-76, 1994.

Anfiloff, V., Polycyclic rifting-an interpretation of gravity and magnetics in the North West Shelf, in: *The North West Shelf, Australia: proceedings of petroleum exploration society of Australia symposium*, edited by P. G., and R. R. Purcell, 443-455, 1988.

Blakely, R. J., *Potential theory in gravity and magnetic applications*, Cambridge University Press, New York, 1995.

Boehme, R. S., Stratigraphic response to Neogene tectonism on the Australian Northwest Shelf, *M.S. thesis*, Louisiana State University, Baton Rouge, USA, 1996.

Bowin, C., G. M. Purdy, C. Johnston, G. Shor, L. Lawver, H. M. S. Hartono, and P. Jezek, arc-continent collision in Banda Sea region, *The American Association of Petroleum Geologist Bulletin*, 64, 868-915, 1980.

Bradley, D. C., and W. S. F. Kidd, Flexural extension of the upper continental crust in collisional foredeeps, *Geological Society of America Bulletin*, 103, 1416-1438, 1991.

Brown, C. D., and R. J. Phillips, Crust-mantle decoupling by flexure of continental lithosphere, *Journal of Geophysical Research*, 105, 13,221-13, 237, 2000.

Burov, E. B., and M. Diament, Flexure of the continental lithosphere with multilayered rheology, *Geophys. J. Int.*, 109, 449-468, 1992.

Burov, E. B., and M. Diament, The effective elastic thickness (EET) of continental lithosphere: What does it really mean?, *Journal of Geophysical Research*, 100, 3905-3927, 1995.

Burov, E. B., and M. Diament, Isostasy, equivalent elastic thickness, and inelastic rheology of continents and oceans, *Geology*, 24, 419-422, 1996.

Chamalaun, F. H., and A. E. Grady, The tectonic development of Timor: a new model and its implications for petroleum exploration, *The Australian Petroleum Exploration Association*, 102-108, 1978.

Charlton, T. R., Post collision extension in arc-continent collision zone, eastern Indonesia, *Geology*, 19, 28-31, 1991.

Charlton, T. R., Back-thrusting on the BIRPS deep seismic reflection profiles, Banda Arc, Indonesia, a response to changing slab inclination, *Journal of the Geological Society*, *London*, 154, 169-172, 1997.

Chemenda, A. I., J-P, Burg, M. Mattauer, Evolutionary model of the Himalaya-Tibet system: Geopoem based on new modelling, geological and geophysical data, *Earth and Planetary Science Letters*, 174, 397-409, 2000.

Curry, J. S., J. M. Lorenzo, and Geoff W. O' Brien, Polarity of continent-island arc collision since late Miocene: Timor Sea, N. W. Shelf, *The Australia, American Association of Petroleum Geologists*, A35, 2000.

DeMets, C., R. G. Gordon, D. F. Argus, and S. Stein, Effects of recent revisions to the geomagnetic reversal time scale on estimates of current plate motions, *Geophysical Research Letters*, 21, 2191-2194, 1994.

DeSmet, M. E. M., A. R. Fortuin, S. R. Troelstra, L. J. Van Marle, M. Karmini, S. Tjokrospoetro, and S. Hadiwasastra, Detection of collision-related vertical movements in the outer Banda Arc (Timor, Indonesia), using micropaleontological data, *Journal of Southeast Asian Earth Sciences*, 4, 337-356, 1990.

Engdahl, E.R., R. D. van der Hilst, and R. Buland, Global teleseismic earthquake relocation with improved travel times and procedures for depth determination, *Bull. Seism. Soc. Am*, 88, 722-743, 1998.

Garcia-Castellanos, D., M. Torne, and M. Fernandez, Slab pull effects from flexural analysis of the Tonga and Kermadec trenches (Pacific Plate), *Geophys. J Int.*, 141, 479-484, 2000.

Genrich, J. F., Y. Bock, R. McCaffrey, E. Calais, C. W. Stevens, and C. Subarya, Accretion of the southern Banda Arc to the Australian plate margin determined by global positioning system measurements, *Tectonics*, 15, 288-295, 1996.

Hamilton, W., Tectonics of the Indonesian region, US Govt. Printing Office, 1974.

Harris, R. A., Temporal distribution of strain in the active Banda orogen: A Reconciliation of Rival Hypotheses, *Journal of Southeast Asian Earth Sciences*, 6, 373-386, 1991.

Johnston, C. R., and C. O. Bowin, Crustal Reactions Resulting from the Mid-Pliocene to Recent Continent-island arc Collision in the Timor Region, *Australian Bureau of Mineral Resources Geology and Geophysics Journal*, 6, 223-243, 1981.

Jurdy, D. M., and M. Stefanick, The forces driving the plates: constraints from kinematics and stress observations, Phil. Trans. R. Soc. Lond., 337, 127-139, 1991.

Karig, D. E., A. J. Barber, T. R. Charlton, S. Klemperer, and D. H. Hussong, Nature and distribution of deformation across the Banda Arc-Australian collision zone at Timor, *Geological Society of America Bulletin*, 98, 18-32, 1987.

Karner, G. D., and A. B. Watts, Gravity anomalies and flexure of the lithosphere at mountain ranges, *Journal of Geophysical Research*, 88, 10,449-10,477, 1983.

Karner, G. D., N. W. Driscoll, J. K. Weissel, Response of the lithosphere to in-plane force variations, *Earth and Planetary Science Letters*, 114, 397-416, 1993.

Kennett, B.L.N., E. R. Engdahl, E.R. and R. Buland, Constraints on seismic velocities in the Earth from travel times, *Geophysical Journal International*, 122, 108-124, 1995.

Kreemer, C., W. E. Holt, S. Goes, and R. Govers, Active deformation in eastern Indonesia and the Philippines from GPS and seismicity data, *Journal of Geophysical Research*, 105, 663-680, 2000.

Lorenzo, J. M., G. W. O'Brien, J. Stewart, and K. Tandon, Inelastic yielding and forebulge shape across a modern foreland basin: north west shelf of Australia, Timor Sea, *Geophysical Research Letters*, 9, 1455-1458,1998.

Lyon-Caen, H., and P. Molnar, Constraints on the structure of Himalaya from an analysis of gravity anomalies and a flexural model of lithosphere, *Journal of Geophysical Research, 88*, 8171-8191, 1983.

Lyon-Caen, H., and P. Molnar, Gravity anomalies, flexure of the Indian Plate and the structure, support and evolution of the Himalaya abd Ganga basins, *Tectonics, 4*, 513-538, 1985.

Mackenzie, D., Ocean floor is laid bare by new satellite data, *Science*, 277, 1921, 1997.

McAdoo, D. C., J. G. Caldwell, and D. L. Turcotte, On the elastic-perfectly plastic bending of the lithosphere under generalized loading with application to the Kuril Trench, *Geophys. Journ. Roy. Astron. Soc.*, 54, 11-26, 1978.

McCaffrey, R., Active tectonics of the eastern Sunda and Banda Arcs, *Journal of Geophysical Research*, 93, 15,163-15,182, 1988.

McCaffrey, R., and J. Nabelek, Seismological evidence for shallow thrusting north of the Timor Trough, *Geophys. Journ. Roy. Astron. Soc.*, 85, 365-381, 1986.

McCaffrey, R., P. Molnar, S. W. Roeker, and Y. S. Joyodiwiryo, Microearthquake seismicity and fault plane solutions related to arc-continent collision in the eastern Sunda arc, Indonesia, *Journal of Geophysical Research*, 90, 4511-4528, 1985.

McKenzie, D., and D. Fairhead, Estimates of the effective elastic thickness of the continental lithosphere from Bouguer and free air gravity anomalies, *Journal of Geophysical Research*, 102, 27,523-27,552, 1997.

McNutt, M. K., M. Diament, and M. G. Kogan, Variations of elastic plate thickness at continental thrust belts, *Journal of Geophysical Research*, 93, 8825-8838, 1988.

van der Meulen, M. J., S. J. H. Buiter, J. E. Meulenkamp, and M. J. R Wortel, An early Pliocene uplift of the central Apenninic foredeep and its geodynamic significance, *Tectonics*, 19, 300-313, 2000.

Molnar, P., and D. Gray, Subduction of continental lithosphere: some constraints and uncertainties, *Geology*, 7, 58-62, 1979.

National Geophysical Data Center, ETOPO-5 bathymetry/ topography data, National Oceanic and Atmospheric Administration, US Dept. Commerce, Boulder, Colorado, 1988.

Poudjom Djomani, Y. H., J. D. Fairhead, and W. L. Griffin, The flexural rigidity of Fennoscandia: reflection of the tectonothermal age of the lithospheric mantle, *Earth and Planetary Science Letters*, 174, 139-154, 1999.

Richardson, A., (preprint) Constraints on continent-arc collision in eastern Indonesia, 2001.

Royden, L., and G. D. Karner, Flexure of the continental lithosphere beneath Apennine and Carpathian foredeep basins, *Nature*, 309, 142-144, 1984.

Royden, L., E. Patacca, and P. Scandone, Segmentation and configuration of subducted lithosphere in Italy: an important control on thrust-belt and foredeep basin evolution, *Geology*, 15, 714-717, 1987.

Sandwell, D. T., and W. H. F. Smith, Marine gravity anomaly from Geosat and ERS 1 satellite altimetry, *Journal of Geophysical Research*, 102, 10039-10054, 1997.

Simandjuntak, T. O, and A. J. Barber, Contrasting tectonic styles in the Neogene orogenic belts of Indonesia, in: *Tectonic evolution of southeast Asia*, edited by R. Hall and D. Blundell, *Geological Society Special Publication, 106*, 185-201, 1996.

Simons, F. J., M. T. Zuber, and J. Korenaga, Isostatic response of the Australian lithosphere: Estimation of effective elastic thickness and anisotropy using multitaper spectral analysis, *Journal of Geophysical Research*, 105, 19,163-19,184, 2000.

Smith, W. H. F., and D. T. Sandwell, Global seaflor topography from satellite altimetry and ship depth soundings, *Science*, 277, 1956-1962, 1997

Snyder, D. B., H. Prasetyo, D. J. Blundell, C. J. Pigram, A. J. Barber, A. Richardson, and S. Tjokosaproetro, A dual doubly vergent orogen in the Banda Arc continent-arc collision zone as observed on deep seismic reflection profiles, *Tectonics*, 15, 34-53, 1996.

Sopaheluwaken, J., H. Helmers, S. Tjokrosapoetro, and E. Surya Nila, Medium pressure metamorphism with inverted thermal gradient associated with ophiolite nappe emplacement, *Netherlands Journal of Sea Research*, 24, 333-343, 1989.

Stagg, H. M. J., Tectonic elements of the north west shelf, Australia, scale 1:2500000, *Australian Geological Survey Organization*, Canberra, 1993.

Stewart, J., and A. B. Watts, Gravity anomalies and spatial variation of flexural rigidity at mountain ranges, *Journal of Geophysical Research*, 102, 5327-5352, 1997.

Storti, F., F. Salvani, and K. McClay, Synchronous and velocity-partitioned thrusting and thrust polarity reversal in experimentally produced, doubly-vergent thrust wedges: Implications for natural orogens, *Tectonics*, 19, 378-396, 2000.

Tandon, K, J. M. Lorenzo, and G. O'Brien, Effective elastic thickness of the northern Australian continental lithosphere subducting beneath the Banda Orogen (Indonesia): Inelastic failure at the start of the continental subduction, *Tectonophysics*, 329, 39-60, 2000.

Turcotte, D. L., and G. Schubert, *Geodynamics: Application of Continuum Physics to Geological Problems*, John Wiley & Sons, New York, 1982.

Uyeda, S., and H. Kanamori, Backarc opening and the mode of subduction, *Journal of Geophysical Research*, 84, 1049-1061, 1979.

Veevers, J. J., Sedimentary sequences of the Timor Trough, Timor, and the Sahul shelf, in: *Initial reports of the Deep Sea Drilling Project*, Edited by J. J. Veevers, J. R. Heirtzler, and et al., 27, 567-569, 1974.

Veevers, J. J., J. R. Heirtzler, and et al., *Initial reports of the Deep Sea Drilling Project*, 27, 1060 pp, 1974.

Veevers, J. J., D. A. Falvey, and S. Robins, Timor Trough and Australia: Facies show topographic wave migrated 80 km during the past 3 my., *Tectonophysics*, 45, 217-227, 1978.

Warris, B. J., Plate tectonics and the evolution of the Timor Sea, northwest Australia, *The Australian Petroleum Exploration Association*, 13-18, 1973.

Watts, A. B., S. H. Lamb, J. D. Fairhead, and J. F. Dewey, Lithospheric flexure and bending of the central Andes, *Earth and Planetary Science Letters*, 134, 9-21, 1995.

Watts, A. B., The strength of the earth's crust, *Marine Technology Society Journal*, 17, 5-17, 1983.

Wessel, P., Analytical Solutions for 3D flexural deformation semi-infinite elastic plates, *Geophysical Journal International*, 124, 907-918, 1996.

Wessel, P., and W. H. F. Smith, New version of generic mapping tools, *Eos. Trans. AGU Electron. Suppl.*, Aug. 15, 1995. (Available at http://www.agu.org/eos-elec/ 9514e.html)

Widiyantoro, S., and R. D. van der Hilst, Structure and evolution of lithospheric slab beneath the Sunda arc, Indonesia, *Science*, 271, 1566-1570, 1996.

Wong a Ton, S. Y. M., and M. J. R. Wortel, Slab detachment in continental collision zones: An analysis of controlling parameters, *Geophysical Research Letters*, 24, 2095-2098, 1997.

Juan M. Lorenzo, Department of Geology and Geophysics, Louisiana State University, Baton Rouge, LA 70803, USA. (juan@geol.lsu.edu)

Geoffrey W. O'Brien, Australian Geological Survey Organization, GPO Box 378, Canberra, Australia, 2601. (gobrien@selenite.agso.gov.au)

Kush Tandon, Department of Geology and Energy Institute, University of Louisiana, Lafayette, LA 70503, USA. (kxt2777@louisiana.edu, kushtandon@yahoo.com)

Sri Widiyantoro, Department of Geophysics and Meteorology, Bandung Institute of Technology, Bandung, Indonesia. (sriwid@geoph.itb.ac.id)