GEOPHYSICAL RESEARCH LETTERS, VOL. 25, NO. 9, PAGES 1455-1458, MAY 1, 1998

# Inelastic yielding and forebulge shape across a modern foreland basin: North West Shelf of Australia, Timor Sea

Juan M. Lorenzo[1], Geoffrey W. O'Brien[2], Jonathan Stewart[3] and Kush Tandon[1]

**Abstract.** The Timor Trough is a modern 'underfilled' foreland basin created by partial subduction of the outer north west continental shelf of Australia beneath Timor Island in the Outer Banda Arc of eastern Indonesia during the Cenozoic. A change of the effective elastic thickness (EET) of the continental foreland lithosphere from ~80 ±20 km to ~25 km over a distance of ~300 km explains (1) the high curvature (~$10^{-7}$ $m^{-1}$) on the outer Trough wall, (2) the low shelf forebulge (~200 m) as measured along a reference base Pliocene unconformity, and (3) observed gravity. An inelastically yielding quartzite–quartz-diorite–dunite continental rheology can explain the EET gradient. New, shallow crustal (<8 km), seismic reflection images indicate that Jurassic basement normal faults are reactivated during bending of the foreland.

## Introduction

Relatively little attention has been paid to the effect of shallow crustal normal faulting e.g., [*Bradley and Kidd*, 1991] and deeper ductile yielding on the shape of continental foreland forebulges. Most previous studies explain the first-order geometry of down-flexing continental basement in ancient, mature foreland basins e.g., [*Karner and Watts*, 1983] and their stratigraphic evolution e.g., [*Coakley and Watts*, 1991] using thin, elastic plate models which predict an upwarp (forebulge) of the order of $10^1$ -$10^2$ km wide at the periphery of basins. The net, effective lithospheric strength from foreland basins worldwide commonly has been expressed as the effective elastic plate thickness (EET) which appears to be bimodally distributed around either weak (10-20 km) or strong (80-90 km) values [*Watts*, 1992]. Most of these studies have assumed a homogeneous lithosphere of constant strength in space and time.

When bending stresses exceed rock strength, the lithosphere can experience brittle and ductile failure (Fig. 1A-B). For an initially strong foreland lithosphere, increased bending toward the orogen can reduce EET by at least half [*Waschbusch and Royden*, 1992]. Permanent deformation within the lithosphere, will modify the stress distribution and change the observed deflection of the yielding basement. Another ~40% reduction may occur if the mantle is decoupled from a failed lower crust [*McNutt et al.*, 1988; *Burov and Diament*, 1995]. As a result of these rheological changes, the foreland basin can become narrower and deeper (Fig. 1A). If the Moho, a surface within the elastic plate, parallels the surficial flexural topography, then forward modeling of the free-air gravity field can be used to verify the first-order geometry of the bending plate at depth.

Few seismic images exist across modern forebulges. Herein, we examine a foreland basin on the outer north west continental shelf of Australia that has recently undergone shallow normal faulting (no more than 8 km) during partial subduction beneath Timor Island (Outer Banda Arc), eastern Indonesia (Fig. 2). The foreland surface is interpreted from new seismic reflection data correlated to commercial well data (Figs. 3 & 4), [*Boehme*, 1996]. Using constraints from gravity data we attempt to explain the shape of a reference foreland surface and presence of normal faults across the outer shelf and Timor Trough by analogy to a yielding, flexing plate.

## Inelastic Yielding and Mechanical Decoupling

We use a composite rheological model for continental lithosphere [*Burov and Diament*, 1995] to determine the geological bounds of sustainable stresses (Fig. 1B) during bending. Brittle faulting occurs if stresses exceed a frictional failure criterion [*Byerlee*, 1978] which depends on confining pressure and the presence of water. Yield strength in the lower crust and lithosphere is predicted by temperature-sensitive ductile flow relations that incorporate strain-dependence and rock type [*Brace and Kohlstedt*, 1980]. We assume that once the yield stress is reached, additional bending leads only to unrecoverable strain. Under these circumstances, curvature can continue to grow concentrating bending in weak yielding zones without the increase in stress expected for a simple elastic plate-- as a result, EET decreases.

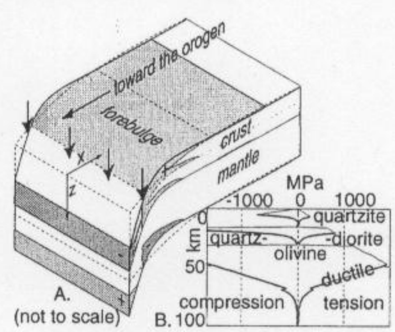


**Figure 1 A.** Schematic block diagram comparing an end-loaded (vertical black arrows) broken plate of constant effective elastic thickness (dashed) versus one plate that has failed inelastically (shaded regions) either under compression (-) or extension (+) within the crust and mantle. Curvature is negative for convex upward geometries. B. Strength (MPa) for an 'intermediate' strength lithosphere. Thick lines mark stresses at or beyond which ductile yielding occurs.

[1]Department of Geology and Geophysics, Louisiana State University Baton Rouge.
[2]Australian Geological Survey, Canberra, Australia.
[3]Department of Earth Sciences, University of Oxford, England.



Paper number 98GL01012.
0094-8534/98/98GL-01012$05.00

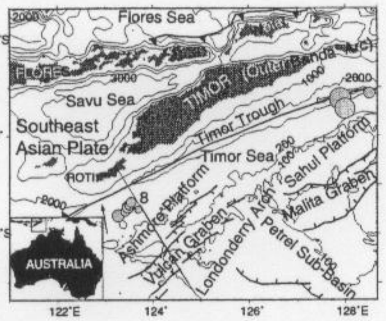


**Figure 2.** Outer North West Shelf of Australia (inset) in collision with Southeast Asia along Timor Trough. Thick line marks seismic track. Thin line spans region seen in Fig. 5. Circles locate modern seafloor fault scarps –size is proportional to vertical throw. Fault line barbs point to buried grabens. Arrow indicates local Australia-Eurasia convergence direction at 76.6 +/-1.1 km/my [*DeMets et al.,* 1994].

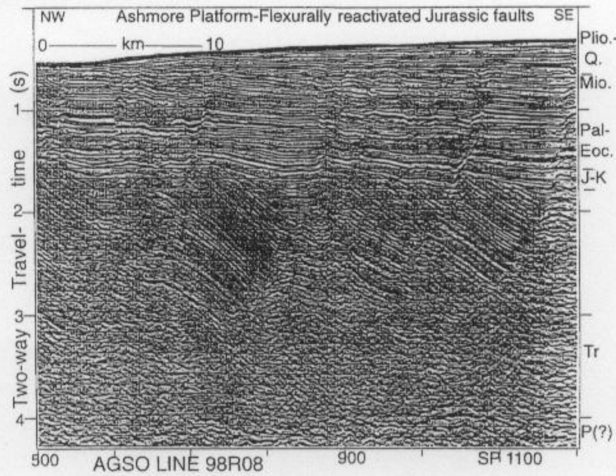


**Figure 4.** Uninterpreted migrated seismic image for Line 8 (Fig.2); abbreviations as in Fig. 3.

[*McNutt,* 1984] shows that at a given position along the plate EET can be obtained from curvature and stored moment. For this study, curvature of the underthrust continent is calculated pointwise as the second derivative of the best-fitting flexure curve through the base Pliocene surface. Within the constraints of the rheological model we estimate the stored moment by integrating bending stresses numerically to 100 km. At this depth, given the assumed strain rate and geothermal gradient below, the mantle has 0 strength (Fig. 1B). Because plate strength is sensitive to yield history, we also track the stress distribution through time [*Waschbusch and Royden,* 1992] assuming that the present best-matched flexural profile has remained constant. Further reduction of EET can occur if during inelastic yielding the lithosphere becomes subdivided into independently flexing sublayers which facilitate additional bending. These decoupled sublayers reduce lithospheric strength approximately to that of the thickest competent unit [*Burov and Diament,* 1995] which in our case is the competent mantle.

## Geological Background

The Timor Sea area is a modern analogue for 'underfilled' foreland basins [*Shanmugan and Lash,* 1982] which form on the cratonward side of thrust belts above subducting plates. On the island of Timor the oldest evidence for collision, dated radiometrically at ~38 Ma (Late Eocene), is recorded as a prograde metamorphic event in metasediments beneath an overthrust ophiolite complex [*Sopaheluwakan et al.,* 1989]. Although the Australian Plate continues to subduct along the Java Trench to the west of Timor, seismicity [*McCaffrey,* 1988] and GPS [*Genrich et al.,* 1995] studies indicate that Australia–Timor convergence has ceased and that Australia – Eurasia convergence is accommodated by backthrusting north of Timor along the Flores and Wetar thrusts. Tensional earthquake events and the absence of compressional events [*Christensen and Ruff,* 1988] support a regional tensional regime in the Timor Sea.

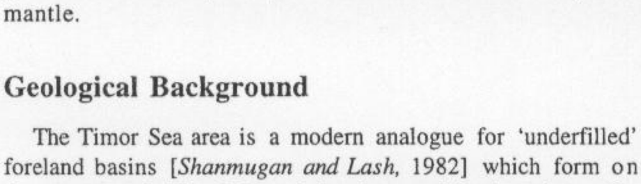


**Figure 3.** Line interpretation of Australian Geological Survey Organisation (AGSO) seismic reflection image. Ages of units are correlated to nearby wells (thick black vertical lines). Age abbreviations: Plio(cene)-Q(Recent), Mio(cene), J(urassic)-K(Cretaceous), Tr(iassic), P(ermian). Depth scale (km) below seafloor and true dips (°) are calculated from well velocity logs.

Since the Paleozoic, the Australian continental shelf has experienced multiple rifting events which created a series of isolated sediment-starved platforms (Londonderry Arch, Ashmore and Sahul Platforms) with intervening depocenters (e.g., Vulcan and Malita Grabens) (Fig. 2). Jurassic-age rift basins trending ENE and NE on the outer shelf were developed at a high angle to an older (Late Devonian–Early Carboniferous) NW-trending fault system (Petrel Sub-Basin) [*Gunn,* 1988]. Potentially, such rifting could render the continental lithosphere weak for extended periods of time (>$10^8$ m.y.) [*Watts,* 1992]. However, seismic refraction measurements [*Bowin et al.,* 1980] indicate that the range of crustal thickness values on the shelf (~31-34 km) is spanned by those determined for the Timor Trough (~27.5-36.5 km) and do not seem to reflect an inherited rift structure that would significantly weaken the lithosphere toward the Trough.

A line interpretation of new seismic data collected to the SE of Timor (Fig. 2) shows many shallow normal faults that were created by reactivation of deeper Jurassic rift-bounding faults. Some faults appear to detach within Cretaceous units (Figs. 3 and 4). Most reactivated faults terminate above, or near, a regional Base Pliocene unconformity [*Boehme,* 1996] (Fig. 3), which constrains the end of the extension to latest Miocene/earliest Pliocene. Characteristically, in seismic images (e.g., Fig. 4) , the latest Miocene faults are steeper than the underlying planar-normal Jurassic faults with which they link, and together produce an apparent listric geometry. Jurassic faults are often traceable into the Triassic section and sole out into a discontinuous high-amplitude "top-Permian" reflector.

The base Pliocene unconformity provides a potential preflexural reference surface for forward modeling EET. Based on present-day convergence rates between Australia and Eurasia (Fig. 2) in the early Pliocene the current shelf/slope transition would have been located ~350 km further south and therefore would not have been significantly affected by flexure at the time of its formation.

## Determining EET

Similar to previous workers, we employ a broken elastic plate analog which has a laterally variable EET embedded in a fluid mantle [*Bodine et al.*, 1981; *Stewart and Watts*, 1997]. We tested separately EET gradients north and south of the Trough axis, and then adjusted the northern limit of the broken plate to improve the fit of the model to both the island topography and the Pliocene reference surface (Fig. 5A). A broken plate agrees with deep seismic reflection images east of Timor [*Richardson and Blundell*, 1996] that show highly disrupted upper continental crust under the center of the orogen. At greater depths, lithospheric break-off has also been proposed to explain a vertical zone of weakness inferred from earthquake fault plane solutions in the Savu Sea [*McCaffrey et al.*, 1985]. The end of the elastic plate does not correspond necessarily to a physical boundary but is where the elastic lithosphere may be effectively decoupled from deeper lithosphere and or where it has begun to lose significant rigidity. The resultant flexural geometry is loaded vertically by the forearc north of the Timor Trough, the Pliocene-Recent sediments south of the Trough, and infilling seawater. The net effect of hidden loads [*Karner and Watts*, 1983] such as residual slab pull forces is approximated by additional vertical end forces.

Figure 5A displays the lateral range of EET values (~25 km to 80 ± 20 km over 300 km) estimated by modeling the base

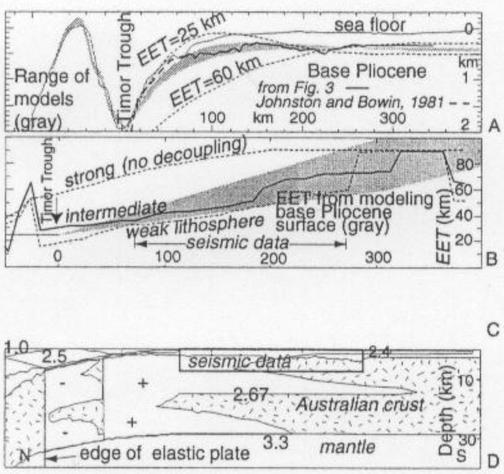


**Figure 5** A. Base Pliocene foreland surface (thick line) along Line 8 (Fig. 2) compared to a range of flexural geometries (gray region) predicted using laterally variable effective elastic thickness (EET). For the same load, dashed lines show flexural shape using a constant EET. B. Estimates for EET based on modeling base Pliocene surface (Fig. 5A) compared to predicted changes in EET caused by inelastic yielding and mechanical decoupling (thick black and dashed lines). All-olivine lithosphere is 'strong'. Decoupling is required for both 'weak' and 'intermediate' lithosphere to match low EET values. Peaks in predicted EET are caused by unbending lithosphere. C. Gravity data fit by preferred density model D. Preferred density (g/cm3) model using a flexed Moho with shape from average modeled estimate in Fig. 5B. Regions predicted to fail in extension (+) and in compression (-) are calculated using an 'intermediate' strength rheological model (Figs. 1A-B, 5B).

**Table 1.** Model Parameter Values

| Rheology | Activation energy (J mol$^{-1}$) | n | A(Pa$^{-n}$s$^{-1}$) |
|---|---|---|---|
| Quartzite | $1.9 \times 10^5$ | 3 | $5.01 \times 10^{-12}$ |
| Quartz-diorite | $2.12 \times 10^5$ | 2.4 | $5.01 \times 10^{-15}$ |
| Olivine-dunite | $5.2 \times 10^5$ | 3 | $7 \times 10^{-14}$ |
| ($\sigma_1$-$\sigma_3 \geq 200$ MPa) | $5.35 \times 10^5$, $\sigma_0$=$8.5 \times 10^3$ MPa, $\varepsilon_0 = 5.7 \times 10^{11}$ s$^{-1}$ | | |
| Young's modulus: | $0.5 - 5.0 \times 10^{10}$ N m$^{-2}$ | | |

Pliocene surface. Constant-EET models only match segments of the base Pliocene surface. For example, in the Trough, an EET = 25 km matches the curvature (~2-6 $\times 10^{-7}$ m$^{-1}$) but a larger EET = 60 km is needed over the shelf. We have considered the effect of in-plane extensional stresses which could decrease the forebulge amplitude, but our models suggest that the necessary force ($10^{14}$ N m$^{-1}$) is greater than expected from slab pull [*Jarrard*, 1986] and would fracture the entire continental shelf [*Bowin et al.*, 1980], in contradiction of observations from the seismic profile (Fig. 3). Resolution of EET estimates decreases with decreasing curvature toward the shelf. Individual EET values can have uncertainties of at least 25% for reasonable rheological parameter values during flexure modeling [*Burov and Diament*, 1995]. All models shown here require a vertical end force ~ $0.5 \times 10^{12}$ N m$^{-1}$ to fit the observed flexural profiles.

To examine whether expected changes in EET caused by inelastic yielding and decoupling can explain interpreted EET lateral gradients we employ two end-member and one intermediate-strength continental rheologies (Fig. 5B) (Table 1, e.g., [*Burov and Diament*, 1995]). Our preferred rheological model which best predicts the EET gradient trend is of 'intermediate' strength (assumed strain rate=$5 \times 10^{-15}$ s$^{-1}$). Conservatively, we assume that the geothermal gradient is at least equal to that of the thermally older Precambrian Central Australian Shield, which has been theoretically estimated at ~13°C/km [*Lachenbruch and Sass*, 1977]. We fix the pore-to-lithostatic pressure ratio: $\lambda$=0. Our 'strong' lithosphere uses an all-olivine-dunite rheology for the crust and mantle, a low geothermal gradient (10°C/km) and an assumed strain rate of $10^{-14}$ s$^{-1}$. Our 'weak' case uses a quartzite crust, a high crustal geothermal gradient (20°C/km), a high $\lambda$ (0.7) and a lower strain rate ($10^{-15}$ s$^{-1}$). In this case, high pore pressures facilitate brittle failure [*Brace and Kohlstedt*, 1980]. Figure 5D shows that decoupling is needed for the weak case, to match the estimated EET gradient and values. The homogeneous 'strong' case does not fail internally into necessary independent layers.

Although various models can be used to fit the gravity field, a flexural shape for the seafloor and Moho most simply explains the observed free-air anomaly (Fig. 5C). In the model, the underthrust plate dips north ~ 3° under the island load and ends ~70 km north of the Timor Trough, just north of the island center (Fig. 5D). Greater dips (>30°) are interpreted from seismic images east of Timor [*Richardson and Blundell*, 1996] but this geometry can not be reconciled with gravity field data. Poorer fits to the data occur over the island load where the geometry of the structure is unconstrained seismically.

## Discussion and Conclusions

In the Timor Sea, the shape of the base Pliocene surface may be explained by a lateral reduction in the EET through inelastic yielding and mechanical decoupling within the crust and between the crust and mantle. In contrast, EET gradients across foreland basins have generally been attributed to inherited compositional and thermal differences [*Wu*, 1991] or inherited long-term weakening of the lithosphere by previous rifting episodes [*Watts*, 1992]. Horizontal gradients in EET across the Australian continental lithosphere (0.12-0.25) are either comparable to other foreland basins (Alberta foreland basin (0.16-0.25) [*Wu*, 1991] ; Romanian Carpathians (~0.23) [*Stewart and Watts*, 1997] ) or slightly greater (Brazilian Andes (~0.10), [*Stewart and Watts*, 1997] ).

Our results rely on the accuracy of the known stratigraphic history of the foreland reference surface. For example, erosion can lower the forebulge amplitude (Fig. 5B) and lead to overestimation of EET. Undocumented erosion in the Timor Trough could also improve the fit to our flexure model. Although we can not totally quantify erosion of the forebulge area, the Miocene sediment units underlying the foreland surface increase in apparent thickness toward the Trough (Fig. 3) and thus do not support extensive erosion in the forebulge region. By contrast, in many ancient systems forebulge uplift has usually been inferred indirectly from the presence and nature of erosional surfaces, or interpreted changes in sedimentation rate, e.g., [*Shanmugan and Lash*, 1982]. Moreover in these cases, the influence of ancestral topography on the position and shape of the forebulge, is critical and remains unclear.

Our results may have implications in oceanic zones of convergence where the commonly assumed rheology is homogeneous, olivine-rich and strong so that the degree of inelastic yielding should be smaller [*Brace and Kohlstedt*, 1980]. However, plate curvatures can be large (e.g. > $5 \times 10^{-7}$ m$^{-1}$ [*Judge and McNutt*, 1991]) and inelastic yielding can also suppress forebulge height by a factor of 2 or 3 [*McAdoo et al.*, 1978]. *Kao and Chen* [1996] correlate large ($M_w$ or $M_s \geq 7.6$) normal fault earthquakes fracturing the entire brittle lithosphere to partially collapsed forebulges.

**Acknowledgments.** We thank P. DeCelles, M. Steckler, J. Nunn, P. Wessel, and reviewers B. Coakley and E. Silver for the many suggestions that markedly improved this paper. This research was supported by an Oak Ridge Associated Universities Award, and an LSU Summer Stipend Research Award. Acknowledgment is made to the donors of the Petroleum Research Fund, administered by the American Chemical Society for partial support of this research. Maps are generated using GMT software by P. Wessel and W. Smith.

## References

Bodine, J.H., M.S. Steckler, and A.B. Watts, Observations of flexure and the rheology of the oceanic lithosphere, *J. Geophys. Res., 86*, 3695-3707, 1981.

Boehme, R.S., Stratigraphic response to Neogene tectonism on the Australian Northwest Shelf, M.Sc. thesis, 71 pp., Louisiana State University, 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, *AAPG Bull., 64*, 868-915, 1980.

Brace, W.F., and D.L. Kohlstedt, Limits on lithospheric stress imposed by laboratory experiments, *J. Geophys. Res., 85*, 6248-6252, 1980.

Bradley, D.C., and W.S.F. Kidd, Flexural extension of the upper continental crust in collisional foredeeps, *Geol. Soc. Amer. Bull., 103*, 1416-1438, 1991.

Burov, E.B., and M. Diament, The effective elastic thickness (Te) of continental lithosphere: What does it really mean?, *J. Geophys. Res., 100*, 3905-3927, 1995.

Byerlee, J.D., Friction of rocks, *Pure Appl. Geophys., 116*, 615-626, 1978.

Christensen, D.H., and L.J. Ruff, Seismic coupling and outer rise earthquakes, *J. Geophys. Res., 93*, 13421-13444, 1988.

Coakley, B.J., and A.B. Watts, Tectonic controls on the development of unconformities: The North Slope, Alaska, *Tecton., 10*, 101-130, 1991.

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

Genrich, J., Y. Bock, R. McCaffrey, E. Calais, C. Stevens, J. Rais, C. Subarya, and S.S.O. Puntodewo, Kinematics of the eastern Indonesia Island Arc estimated by global positioning system measurements, *EOS, Trans., AGU, 75*, 162, 1995.

Gunn, P.J., Bonaparte rift basin: effects of axial doming and crustal spreading, *Explor. Geophys., 19*, 83-87, 1988.

Jarrard, R.D., Relation among subduction parameters, *Rev. Geophys., 24*, 217-284, 1986.

Judge, A.V., and M.K. McNutt, The relationship between plate curvature and elastic plate thickness: A study of the Peru-Chile Trench, *J. Geophys. Res., 96*, 16,625-16,639, 1991.

Kao, H., and W.-P. Chen, Seismicity in the outer rise-forearc region and configuration of the subducting lithosphere with special reference to the Japan Trench, *J. Geophys. Res., 101*, 27811-27831, 1996.

Karner, G., and A. Watts, Gravity anomalies and flexure of the lithosphere at mountain ranges, *J. Geophys. Res., 88*, 10449-10477, 1983.

Lachenbruch, A.H., and J.H. Sass, Models of an extending lithosphere and heat flow in the Basin and Range Province, in *Cenozoic Tectonics and Regional Geophysics of the Western Cordillera*, edited by R.B. Smith, and G.P. Eaton, pp. 209-250, Boulder, Colorado, 1977.

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. J. Roy. Astron. Soc., 54*, 11-26, 1978.

McCaffrey, R., Active tectonics of the eastern Sunda and Banda arcs, *Journ. Geophys. Res., 93*, 15163-15182, 1988.

McCaffrey, R., P. Molnar, S.W. Roecker, and Y.S. Joyodiwiryo, Microearthquake seismicity and fault plane solutions related to arc-continent collision in the eastern Sunda arc, Indonesia, *J. Geophys. Res., 90*, 4511-4528, 1985.

McNutt, M., M. Diament, and M.G. Kogan, Variations of elastic plate thickness at continental thrust belts, *J. Geophys. Res., 93*, 8825-8838, 1988.

McNutt, M.K., Lithospheric flexure and thermal anomalies, *J. Geophys. Res., 89*, 11180-11194, 1984.

Richardson, A.N., and D.J. Blundell, Continental collision in the Banda arc, in *Tectonic Evolution of Southeast Asia*, Eds. R. Hall, and D. Blundell, pp. 47-60, Geol. Soc. Lond., London, 1996.

Shanmugan, G., and G.G. Lash, Analogous tectonic evolution of the Ordovician foredeeps, southern and central Appalachians, *Geol., 10*, 562-566, 1982.

Sopaheluwakan, J., H. Helmers, S. Tjokrosapoetro, and E. Surya Nila, Medium pressure metamorphism with inverted thermal gradient associated with ophiolite nappe emplacement, *Neth. J. Sea Res., 24*, 333-343, 1989.

Stewart, J., and A.B. Watts, Gravity anomalies and spatial variations of flexural rigidity at mountain ranges, *J. Geophys. Res., 102*, 5327-5352, 1997.

Waschbusch, P.J., and L.H. Royden, Spatial and temporal evolution of foredeep basins: Lateral strength variations and inelastic yielding in continental lithosphere, *Bas. Res., 4*, 179-195, 1992.

Watts, A.B., The effective elastic thickness of the lithosphere and the evolution of foreland basins, *Bas. Res., 4*, 169-178, 1992.

Wu, P., Flexure of lithosphere beneath the Alberta Foreland Basin: Evidence of an eastward stiffening continental lithosphere, *Geophys. Res. Lett., 18*, 451-454, 1991.

J. Lorenzo (e-mail: juan@tellus.geol.lsu.edu) and K. Tandon, Department of Geology and Geophysics, Louisiana State University, Baton Rouge, LA 70803-4101.

J. Stewart (e-mail: Jonathan.Stewart@earth.ox.ac.uk), Department of Earth Sciences, Oxford University, Oxford OX1 3PR, England

G.W. O'Brien (e-mail: gobrien@selenite.agso.gov.au), Australian Geological Survey, GPO Box 378, Canberra, Australia, 2601.

(Received: October 27, 1997; revised: March 12, 1998;
accepted: March 17, 1998.)