THE GEOMETRY OF BACK ARC THRUSTING ALONG THE EASTERN SUNDA ARC, INDONESIA:
CONSTRAINTS FROM EARTHQUAKE AND GRAVITY DATA

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Abstract. The Flores earthquake of December 23, 1978, represents the first seismological evidence for active back arc thrusting behind the eastern Sunda Arc. This m_b = 5.8 earthquake occurred north of Flores Island, where seismic reflection profiling has revealed back arc thrusting, interpreted to be a reaction to compression of the arc following collision with the Australian continent. We investigate the source mechanism and depth of the Flores earthquake by inversion of long-period P waveforms and relocate the earthquake's epicenter incorporating arrival time data from local stations. We find that this event occurred beneath the accretionary prism south of the back arc thrust zone at a depth of 11.5 km (7 km below the seafloor). The best fitting fault plane solution is consistent with active southward thrusting of the floor of the Flores Basin beneath the volcanic arc along a 30° dipping fault plane. The morphology of the thrust zone and the free-air gravity profile over the Flores Basin in the epicentral region resemble those of oceanic trenches, and the observed gravity field is best interpreted if the crust of the Flores Basin dips smoothly below the accreted wedge at an average angle of about 6°. The position of the earthquake hypocenter south of the Flores Thrust and the fault plane solution suggest that the 1978 earthquake is analogous to thrust events at subduction zones and represents slip between the subducting and overriding plates. The Flores Thrust is thus the surface expression of a deep-seated thrust zone and may represent the initial stage of polarity reversal of the eastern Sunda Arc.

Introduction

One of the consequences of plate motions is the eventual involvement of continents and their fragments in convergence zones. A few areas of the world are thought to derive their structural complexity from the attempted subduction of continental crust, but the details of the process are far from obvious. Involvement of these nonsubductable terrains in a convergent boundary may not always result in a large-scale deformation as in the continent-continent collision between India and Asia, but is nevertheless likely to trigger changes in the subduction regime.

McKenzie [1969] suggested that attempted subduction of continental lithosphere beneath an oceanic island arc might result in reversal of subduction polarity because the buoyancy of the continental crust would prevent it from being

subducted. A likely place for this to have happened in the past is northern New Guinea during the Tertiary [Dewey and Bird, 1970; Johnson and Molnar, 1972; Hamilton, 1979], but reversal is not required [Johnson and Molnar, 1972; Johnson and Jaques, 1980]. Reversal may also be underway behind the Sangihe Arc in the northern Molucca Sea [Silver and Moore, 1978; Stewart and Cohn, 1979; Moore and Silver, 1983] following a collision between island arcs. Arc polarity reversal is undoubtedly difficult to identify and prove from the geologic record alone, and thus it is important to study any modern example of where it may possibly be occurring.

Probably the best area to study the incipient subduction of continental material and back arc thrusting is the eastern Sunda Arc, where Australia is presently impinging on the subduction zone. South of Timor and around the great bend in the Banda Arc to the north, continental crust of Australia is found seaward of the Timor Trough (Figure 1) and extends arcward to beneath the trough itself [Jacobson et al., 1979]. Reflection profiles crossing the Australian shelf and the Timor Trough show typical subduction zone morphology and clearly indicate thrusting of Australia beneath the deformed outer arc ridge (which includes Timor) south of the volcanic islands [Jacobson et al., 1979; von der Borch, 1979; Hamilton, 1979; Bowin et al., 1980]. West of Sumba, oceanic lithosphere of the Indian Ocean plate is being consumed at the Java Trench. The presence of the Scott Plateau, possibly a subsided block of the Australian shelf, and Sumba, possibly a continental fragment caught in the fore arc, adds a great deal of complexity to the convergent margin in the transition zone between the Java Trench and the Timor Trough.

Because of the young age of Australia's collision with the eastern Sunda Arc, it has thus far caused only small morphological changes in the subduction system. The available evidence indicates that back arc thrusting has been initiated [Hamilton, 1979; Usna et al., 1979; Silver et al., 1983] that may eventually result in either reversal of the subduction polarity, thus removing the back arc basin from the surface of the earth, or in crustal thickening by imbricate thrusting, thereby producing a belt of ophiolites. The present mode of deformation appears to favor the former scenario.

North of the Sunda island arc from western Sumbawa to east of Wetar (except for a small stretch between east Flores and Alor), seismic reflection profiling has revealed thrusting at the base of the slope of the volcanic pile (Figure 1) [Silver et al., 1983]. The sense of thrusting is such that oceanic crust of the Flores Basin and the Banda Sea is being over-
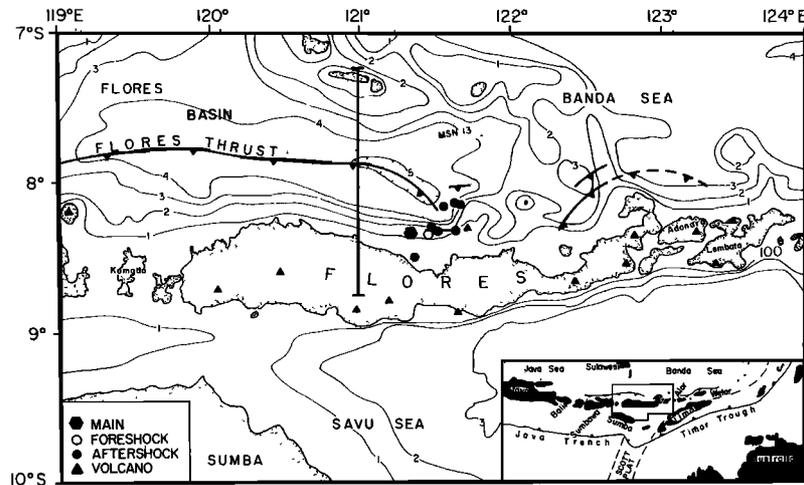
Fig. 1. Location map (insert) and detailed map of the Flores region. Thrust faults
are indicated by solid lines with barbs on the overriding plate, and bathymetric
contours are in kilometers [from Silver et al., 1983]. The dashed line in the inset
shows the inferred northwestern limit of Australian continental crust (and the Scott
Plateau) beneath the outer arc ridge [from Hamilton, 1979]. The large hexagonal symbol
shows the calculated epicenter of the 1978 earthquake. The north trending heavy line
is the position of the profile of Figures 5 and 6. The dotted line north of the Flores
thrust is the location of seismic refraction profile MSN-13.

ridden by the volcanic islands. Silver et al.
[1983] estimate the amount of convergence between
the Flores Basin and the island arc in the Flores
region at 30-60 km based on the volume of
material accreted to the toe of the volcanic arc.
This length is consistent with the amount of
northward indentation of this portion of the arc
with respect to those portions behind which
thrusting is not observed. There is, however, no
evidence that this thrust zone is associated with
a southward dipping zone of intermediate-depth
earthquakes [Cardwell and Isacks, 1978].

In this paper we discuss an m_b = 5.8 earth-
quake from the thrust zone north of Flores; its
fault plane solution constitutes the first
seismological evidence to support interpretations
that active back arc thrusting occurs behind the
Sunda Arc. Also, simple models for the structure
of the back arc thrust zone that satisfy gravity
data and published seismic refraction and
reflection profiles are presented. These data
suggest that the crust (probably oceanic) of the
Flores Basin either dips beneath the small back
arc accretionary prism at an angle of about 6° or
thickens considerably under the southern Flores
Basin. The presence of a large negative -1.10
mm/s^2 (-110 mGal) free-air anomaly over the
accreted wedge 30 km south of the bathymetrically
deepest part of the Flores Basin signifies that
the crust is out of isostatic equilibrium and
suggests the underthrusting plate is flexed down
as in more developed subduction zones. The
position of the earthquake hypocenter south of
the Flores Thrust and the fault plane solution
indicate that the 1978 earthquake was due to
southward underthrusting of the Flores Basin
crust beneath the overriding island arc.

The Earthquake of December 23, 1978

Location

The earthquake of December 23, 1978, together
with one foreshock and seven aftershocks, was

relocated using arrival time data from the
regional network of Indonesia and southeast Asia
[Hodgson, 1980], from a local network operating
on Sulawesi [McCaffrey and Sutardjo, 1982], and
from International Seismological Centre (ISC)
bulletins. We repicked arrival times for all
regional and many of the World-Wide Standardized
Seismograph Network (WWSSN) stations. The main
shock was located at latitude 8.33°S, longitude
121.34°E (Figure 1) and origin time 0510:50.
This epicenter is only 8 km southwest of the ISC
reported epicenter. The earthquake's depth was
fixed at 11 km in these calculations as deter-
mined from the body wave analysis discussed
below. The depth reported by the ISC was 7 km
based on P wave arrival times and 31 km from pP-P
times.

Our calculated location of the event of
December 23 is about 30 km south of the eastern
end of the Flores Thrust (Figure 1). Since the
regional structure is oriented east-west, the
uncertainty in the location due to lateral varia-
tions in the velocity structure is probably
greater in the north-south direction. Clear P
arrivals at stations in central Sumba, western
Timor, and southern Sulawesi, however, give
strong control on the epicenter. The absence of
reported damage on Flores and the presence of
strong water reverberations in the P wave seismo-
grams (discussed below) indicate that the earth-
quake occurred beneath the sea and not beneath
Flores Island. Furthermore, the period of the
water reverberations requires a water depth of
about 4.5 km, which implies that the epicenter
was at least 15 km to the north or northwest of
that determined from the arrival time data
(Figure 1).

The foreshock and aftershocks were relocated
relative to the main shock by the method of joint
hypocenter determination [Dewey, 1971]. The
depths of these events were fixed at 15 km for
lack of independent depth information, but the
choice of any depth above 50 km for these events

TABLE 1.   Source Parameters Determined From the Inversion of P Waves Using Different Source
Parameterizations and Choices for the Error Function[*]

Moment Tensor

Error Function 1

Normalized moment-tensor components:

Principal axes:

				Eigenvalue	Azimuth	Plunge
M_{xx} = -0.51 ±0.03[†]	M_{xy} = -0.11 ±0.07		P	-0.65 ±0.04	9 ±7°	17 ±1°
M_{yy} = -0.10 ±0.03	M_{xz} = -0.40 ±0.02		T	0.75 ±0.02	156 ±5°	70 ±1°
M_{zz} = 0.61 ±0.01	M_{yz} = 0.09 ±0.02		B	-0.10 ±0.04	276 ±6°	10 ±4°

Moment-tensor norm = 2.61 ±0.30 ×10^{18} N m   Depth = 11.5 ±0.3 km
Decomposition:[‡] Double couple = 87%   Linear vector dipole = 13%
Amplitudes of the source time function elements:[§]
       0.23 ±0.01, 0.37 ±0.03, 0.29 ±0.02, 0.13 ±0.01, -0.02 ±0.02

Constrained Double Couple

Error Function	Depth (km)	Scalar moment (10^{18} N m)	Strike (deg)	Dip (deg)	Rake (deg)	Amplitudes of source time function elements
1	11.4 ±0.4	1.83 ±0.08	114 ±9	30 ±2	114 ±6	0.24 0.40 0.26 0.14 -0.04 (±0.02)
2	10.1 ±0.3	1.69 ±0.10	117 ±6	32 ±1	115 ±3	0.34 0.48 0.19 0.06 -0.07 (±0.02)

[*] The two choices for the error function are discussed in the text.
[†] All quoted uncertainties represent one standard deviation.
[‡] The moment tensor is constrained to be purely deviatoric.
[§] The duration of each source time function element is 2.0 s.

has little effect on the epicentral location.
The foreshock and aftershocks appear to have been
concentrated to the northeast of the main shock,
closer to the eastern termination of the Flores
Thrust (Figure 1) as mapped by reflection pro-
filing [Silver et al., 1983]. The apparent north-
east trend of the epicenters may be ficticious
since the 95% confidence ellipses for most of the
relocations are aligned in that direction.

P wave Analysis

Long-period P waves from the vertical compo-
nent of 10 WWSSN stations were inverted simulta-
neously for the orientation of the double-couple
source, the centroid depth and the far-field
source time function. The source time function
was parameterized by a series of box-car ele-
ments, each 2.0 s in duration (this parameter-
ization is identical to that by Langston [1981]).
Data were inverted with two choices for the error
function that was minimized in a least squares
sense: (1) o_i - s_i and
(2) $o_i/(\Sigma o_j^2)^{1/2}$ - $s_i/(\Sigma s_j^2)^{1/2}$, where o_i are the
amplitudes of the digitized observed waveforms
and s_i are the corresponding synthetic seismogram
amplitudes. The summations with respect to j are
made over the number of data points at a given
station. The first error function uses inform-
ation contained in the amplitude variation
between the stations, while the second is sensi-
tive only to the shapes of the seismograms. When
the second error function is used, the seismic
moment is determined by matching the original
unscaled data in the least squares sense with the
synthetic seismograms determined during the final
iteration. Details of these procedures are de-

scribed by Nábělek [1984] and are similar to
those used by Langston [1981] and Ward [1980],
the main difference being that the source depth
is also determined in our inversion procedure.
Observed seismograms were digitized at half-
second intervals and filtered with a highpass
filter with cutoff at 0.017 Hz in order to remove
the low-frequency drift that was observed at some
stations. The first 60 points (30 s) from the
onset of the direct P wave were used for the
analysis. The onset of the P arrival was deter-
mined from the short-period seismograms but was
allowed to vary by up to 1.5 s if improved cross
correlation between the observed and synthetic
waveforms could be found. Because the inversion
using a full moment tensor source yielded an
almost pure double couple mechanism, the double
couple constraint was subsequently applied. The
solutions using either the moment tensor or
constrained double couple source, and either
choice of the error function are essentially
identical (Table 1). The estimated parameters
for the source centroid are depth 10-12 km,
strike 105-120°, dip 28-32°, and rake 105-120°.
The largest contribution to the total seismic
moment comes from the first four elements of the
source time function; the rupture duration was
therefore about 8 s. The seismic moment is
1.5-2.2 ×10^{18} N m, the exact value being sensi-
tive to the estimate of the source depth.
Estimates of the uncertainties above are based on
formal variances for a given model and on varia-
tions in solutions using different error
functions and different acceptable crustal
models; they are therefore larger than the formal
uncertainties given in Table 1. The north dip-
ping nodal plane is clearly the better resolved
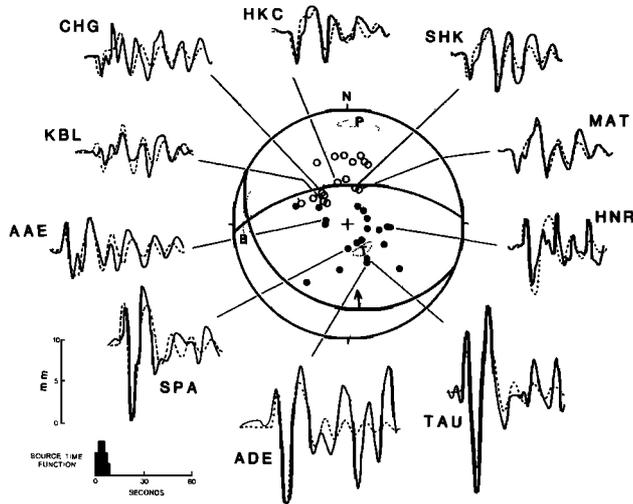
Fig. 2. Observed (solid lines) and synthetic
(dashed lines) long-period P waves and the best
fit fault plane solution for the 1978 Flores
earthquake. The bars on the time scale
represent the source function in 2.0-s
intervals. The amplitude scale for the
seismograms is in millimeters at a magnification
of 3000 and epicentral distance of 40°. Also
shown is a lower hemisphere plot of first motion
polarities. Solid circles represent
compressional first motions, and open circles
are dilatations. The dotted lines enclose the
possible orientations of the P, T, and B axes as
labeled. The arrow shows the horizontal
projection of the slip vector.

of the two. An attempt was made to identify the
actual fault plane from the waveform distortion
that would be expected from rupture propagating
along a finite fault plane, but the results were
inconclusive. Other evidence, discussed below,
suggests that the south dipping nodal plane is
the fault plane.

Observed and synthetic seismograms for the
best fitting fault plane solution are shown in
Figure 2. The observed and theoretical seismo-
grams have been corrected to equal distance (40°)
and instrument magnification (×3000). Also shown
are first motion polarities for P waves that were
read for many more stations, and the uncertain-
ties in the P, T, and B axes. The synthetic
seismograms were generated using the source crus-
tal structure shown in Table 2. This structure
is based on the seismic refraction profile MSN-13
(Figure 1) of Curray et al. [1977]. A value of
1 s was used for t^* (the ratio of travel time to
average Q) to account for anelastic attenuation
in the earth.

Although the synthetic seismograms at the
northern stations appear to satisfy the dilata-
tional first motions observed on the short-period
instruments (a dilatational direct P is clearly
visible on the short-period record from MAT), the
synthetic direct P arrivals at these stations are
in fact compressional. The first discernible
pulse in the long-period records at these sta-
tions corresponds to a strong dilatational pP (P
wave reflection from the sea floor), which
because of the shallow focus of this event,
obscures the direct P arrival [Tréhu et al.,

1981]. The discrepancy between the short-period
first motion data and those predicted by our
model is probably real. Short-period first
motions are indicative of the source orientation
at the nucleation point of the rupture, which may
differ from the average orientation of the fault
plane. The long-period data, however, are indi-
cative of the average orientation. Some varia-
bility in the source orientation is to be
expected considering the long duration of rupture
(8 s), hence a fault length of at least 20 km.

A change in the direction of slip could also
be responsible for the small oscillation observed
in the first few seconds of the long-period
waveform at KBL (Figure 2) that cannot be
accounted for with our point source model.
Alternatively, the precursor at KBL may be ex-
plained by multipath ray propagation through the
high-velocity slab of the subducting Indian Ocean
plate that dips to the north beneath the epi-
central region because the paths of the rays
arriving at this station parallel the strike of
the slab.

In order to investigate the sensitivity of the
waveforms to the source depth, we performed the
inversion at fixed 5-km increments above and
below the best fitting depth (Figure 3). In
terms of the rms residual, the source depth at
11.5 km represents about a 10% improvement over
the solutions at assumed source depths both 5 km
shallower and deeper.

The complexity of the latter part of the
waveforms is mainly due to reverberations within
the water column. The period of these reverber-
ations provides strong constraint on the depth of
the water above the hypocenter, which in turn,
provides a constraint on the epicentral location
when used with bathymetric charts. Figure 4
shows seismograms generated for three water
depths compared to those observed at stations AAE
and MAT. The best fit to the period of the
observed reverberations is at a water depth of
4.5 km; thus this depth was used in calculating
all seismograms and implies a more northerly
location for the epicenter than that determined
from travel time data (Figure 1).

Structure in the Epicentral Region

The Flores earthquake of December 23, 1978,
occurred in the region north of Flores, where
there are large negative free-air gravity anoma-
lies over a small but distinct trench and its
associated deformed accretionary prism. The

TABLE 2. Crustal Structure Used in the
Calculation of the Synthetic Seismograms

Thickness (km)	v_p (km/s)	v_s (km/s)	Density (×10³ kg/m³)
Source region			
4.5	1.51	0.00	1.03
2.0	3.20	1.31	2.30
half-space	6.10	3.10	2.90
Receiver region			
half-space	6.00	3.46	2.80

SHK	HNR	SPA	AAE	HKC	Source Depth km	RMS Error (normalized)	Strike deg	Dip deg	Slip deg	M_o $\times 10^{18}$ N m
					6.5	1.14	115	20	125	4.2
					11.5	1.00	114	30	114	1.8
					16.5	1.07	90	28	97	1.2
					21.5	1.28	87	29	93	0.8
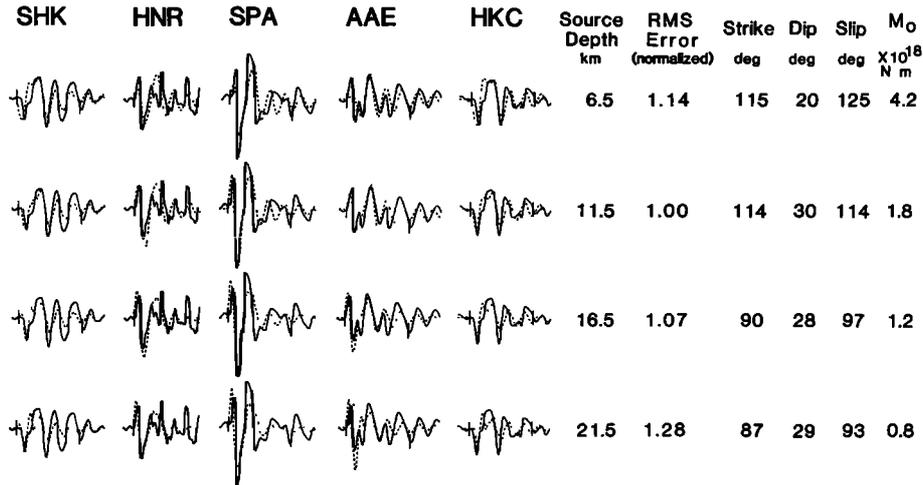
Fig. 3. Results of inverting for source parameters with source depth fixed at 5-km intervals. Solid lines are observed, and dotted lines are calculated seismograms. The depth of 11.5 km was determined by the inversion with the source depth as one of the free parameters and provides the best fitting depth. The rms errors listed are scaled to the minimum value (that for 11.5 km).

strong correlation between the presence of the back arc trench and free-air gravity lows behind the Sunda Arc suggests that the lows result from depression of the crust of the Flores Basin in the process of underthrusting below the arc [Silver et al., 1983]. The gravity lows are most strongly developed north of western Flores over the epicentral region of the earthquake described in this paper. Here, free-air gravity anomalies reach below -1.10 mm/s² (-110 mGal) with the minimum centered over the accretionary prism south of the trench rather than over the bathymetric low of the back arc trench (Figure 5a). In this section we explore crustal structure in the vicinity of the December 23 earthquake by generating simple models satisfying seismic refraction (profile MSN-13 of Curray et al. [1977]), seismic reflection, bathymetric, and free-air gravity data north of Flores [Silver et al., 1983], and Bouguer gravity data on Flores [Geological Research and Development Centre, 1984].

AAE	MAT	Water Depth km
		4.0
		4.5
		5.0

Fig. 4. Observed (solid lines) and synthetic (dotted lines) P wave seismograms generated for water depths of 4.0, 4.5, and 5.0 km. Crustal structure is otherwise as shown in Table 2, and the source parameters are the same as the 11.5 km depth source in Figure 3. A good agreement in period of the water reverberations occurs for a water depths of 4.5 km.

In computing the gravity effect of our two-dimensional models, densities were estimated from seismic velocities obtained from refraction profile MSN-13 and the compressional velocity-density relations of Nafe and Drake [1963] for the sediment layer and of Christensen and Salisbury [1975] for oceanic crustal rocks. We assume that the lowest interface from profile MSN-13 (7.7 km/s at 14-km depth) represents the Moho, but because the density for oceanic crustal rocks of that seismic velocity is close to that of upper mantle rocks [Christensen and Salisbury, 1975], the distinction between rock types is not critical for the purposes of the density model.

The starting model was an isostatically balanced section across the Flores Basin (Figure 5b). The crustal and sediment thicknesses beneath the central Flores Basin were taken from refraction profile MSN-13 that crosses the gravity profile at about km 70 (Figures 1 and 5b). The crust was assumed to be oceanic (i.e., 7 km thick based on MSN-13) where water depths exceed 4 km and to thicken toward the north and south reaching 34 km beneath the coastlines at both edges of the profile. Seafloor topography shows up strongly as short-wavelength gravity anomalies, suggesting a lack of local compensation. The match in amplitudes between observed and calculated anomalies over bathymetric features validates our choice of sediment density. We also included the effect of the subducted Indian Ocean plate, which is likely to have a long-wavelength positive contribution to the gravity field across the profile. The assumed slab geometry (Figure 5c) is based on seismic zone contours from Cardwell and Isacks [1978], a density contrast of 0.05 × 10³ kg/m³ between the slab and asthenosphere [Grow and Bowin, 1975], and a thickness of 50 km. The computed gravity effect of the subducted Indian Ocean plate is 0.54 mm/s² (54 mGal) over Flores Island and decreases to 0.38 mm/s² over the northern end of the profile, constituting a gradient of only 0.001 mm/s² per km (0.1mGal/km). The addition of

A Free-air Gravity

B Isostatically Balanced Model

C Deep Slab Structure

Fig. 5. (a) Free-air gravity profile and line drawing interpretation of seismic reflection profile across the Flores thrust zone [from Silver et al., 1983]. The position of this profile is shown in Figure 1. (b) Comparison of computed gravity profile for an isostatically balanced crustal model (solid curve) with the observed profile (large dots). Note the strong linear trend in the residuals (small dots). Assumed densities (in units of 10^3 kg/m^3) are given for each layer. The isostatically balanced crustal model assumes an average density of 2.8 x10^3 kg/m^3 to a depth of 34 km. Gravity anomalies are free-air over the Flores Basin and Bouguer over Flores Island. (c) Deep slab structure assumed in all calculations and its gravity effect. All horizontal axes are labeled with a common origin.

the positive gravity effect of the slab improves the fit to the observed gravity anomalies over the islands at the ends of the profile which are too positive to be accounted for by the isostatically balanced crustal model alone.

Relative to the isostatically balanced model with the effect of the subducted Indain Ocean plate included, there is a southward decreasing gradient of 0.020 mm/s^2 per km (2.0 mGal/km) over the northern Flores Basin and a total drop in the gravity field of approximately 0.70 mm/s^2 (70

mGal) over the southern Flores Basin (Figure 5b). The residual low relative to the isostatically balanced model is centered over the accretionary wedge and not over the bathymetric low of the trench, suggesting that the crust of the Flores Basin continues to thicken or dip southward below the deformed sediments. The gravity low over the southern Flores Basin requires a mass deficiency beneath the Flores Thrust zone. The requisite crustal thickening can occur either by scraping sediments off the plate as it thrusts coherently beneath the sediment pile along a single, shallow dipping decollement or by a series of imbricate thrust faults involving both the basement and sediment rocks.

The decollement model (Figure 6a) is generated by assuming that the oceanic crust of the Flores Basin is of constant thickness as it thrusts beneath the accretionary wedge. In this case the increase in crustal thickness occurs by down-bowing of the oceanic crust beneath the thrust zone as sediments are accreted to the overlying deformed wedge. In Figure 6a, the crust of the Flores Basin beneath the site of refraction profile MSN-13 agrees in general with those results, and from this point dips to the south at about 6°, reaching a maximum depth of 10.0 km beneath the minimum in the gravity field. The dip angle of 6°, which is likewise the dip angle of the planar sediments south of the Flores Thrust (see the seismic reflection profile in Figure 5a), provides the best fit (within 1°) to the gravity data. The gravity data thus suggest that underthrusting at this dip angle continues tens of kilometers south of the deformation front.

It is interesting that the dip angle determined here is similar to that of many subducting plates at trenches in the western Pacific [Watts and Talwani, 1974]. By analogy we suggest that the Flores Basin lithosphere behaves as an elastic plate while thrusting beneath the island arc and that the large gravity anomalies are due to bending of the plate in response to vertical forces applied to the underthrust portion of the plate. Because the free-air gravity anomalies are negative almost as far south as the north coast of Flores, the load of the crustal material between Flores Island and the Flores Thrust cannot by itself account for the deflection of the Flores Basin lithosphere. Therefore either the underthrusting plate extends as far south as Flores Island or the plate's own gravitational instability aids to pull it into the asthenosphere.

A second mechanism one could envision is the case of a thickened crust perhaps resulting from imbricate thrusting where shortening is taken up along a series of south dipping thrust faults that penetrate both the basement and the sedimentary section. In this case the thickened crust would be comprised of both oceanic crust and sediments. The crustal model in Figure 6b was generated by assuming that the increased thickness of crust beneath the Flores Thrust zone consists of 30% sediment and 70% oceanic crust, in the same proportions as the undisturbed crust that enters the thrust zone. Clearly, this type of model can satisfy the gravity data equally as well as the decollement model. Contrary to the decollement model, closure of the Flores Basin in
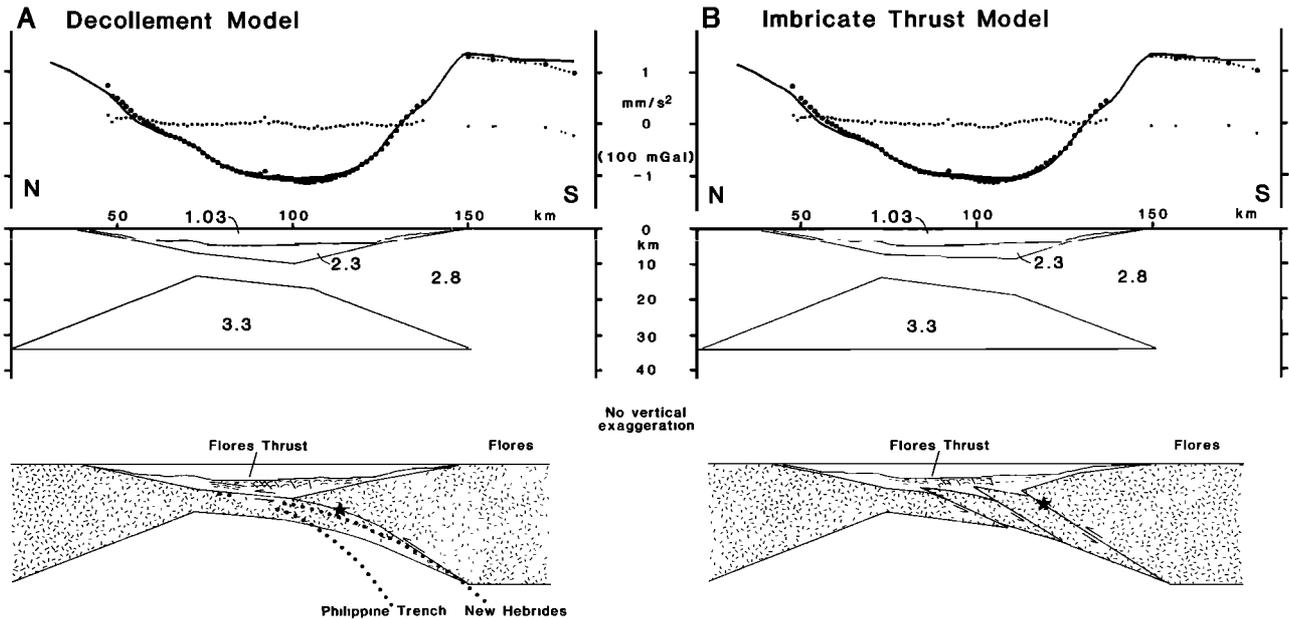
Fig. 6. Alternative models for the structure of the back arc thrust zone. (a) The decollement model where crustal thickening occurs by deflection of the overriding plate and sediment accretion, and (b) the imbricate thrust case in which both oceanic crust and sediments comprise the thickened crust. Presentation of density models is the same as in Figure 5b. At the bottom are cartoon interpretations of the density models. A star shows the inferred position of the 1978 earthquake shifted from its calculated position relative to the crustal models to a position beneath 4.5 km of water as required by the period of reverberations in the waveforms. Dotted lines in the bottom panel of Figure 6a show the projections of the New Hebrides and Philippine Trench seismic zones at the same scale for purposes of comparison. For reasons discussed in the text, the decollement model is preferred.

this fashion would leave its mark in the geologic record as an ophiolite belt sandwiched between two volcanic belts (i.e., Flores and Sulawesi). The increase in the cross-sectional area of the crust shown in Figures 5a and 5b accounts for about 30 km of crustal shortening if the crust entering the thrust zone is 10 km thick initially.

An unlikely alternative is based on the assumption that the north dipping P wave nodal plane of the 1978 earthquake is the fault plane. This implies that Flores Island thrusts beneath the Flores Basin along steeply dipping reverse faults. We consider this scenario highly unlikely because buoyancy forces at the contact between the different crustal types would tend to push the less dense island arc crust out over the oceanic crust when external horizontal compressive forces are applied. Also, the seismic reflection profile in Figure 5a clearly shows planar sediments from the north side dipping southward below the accreted wedge.

Discussion

The Flores earthquake represents the first seismological evidence that active thrusting occurs behind the Sunda Arc north of Flores. We suggest that the south dipping nodal plane is the fault plane because the strike of this plane agrees better with the local strike of the Flores Thrust than does the strike of the steeper plane and the sense of relative displacement (southside

up) on the Flores Thrust is documented in seismic reflection profiles (Figure 5a and Silver et al. [1983]). Furthermore, as shown above, gravity evidence suggests that the Flores Basin crust dips toward and continues under the deformed sediments tens of kilometers south of the thrust zone.

The south dipping nodal plane for the 1978 earthquake dips to the SSW at about 30°. This angle is significantly steeper than the 6° dip of the Flores Basin crust beneath the accreted wedge determined from gravity data and from the dip of planar sediments observed in the reflection profile of Figure 5a. Similarly, along the Philippine Trench, Cardwell et al. [1980] found the dip of the seismic zone defined by earthquake hypocenters and fault planes from shallow thrust events to be 24 ± 10° in the north to 45 ± 15° in the south, although the events examined occurred only 50-100 km landward of the trench. The Philippine Trench is associated with a seismic zone extending to only 150-km depth and is not associated with active volcanism; it is therefore thought to be quite young [Cardwell et al., 1980]. The projections of the Philippine and the New Hebrides seismic zones are shown in Figure 6a for comparison and indicate that the slip angle and position of the Flores earthquake (shown by a star) with respect to the trench are consistent with the subduction interpretation within the limits imposed by observations in other parts of the world.

Silver et al. [1983] discuss the relative

importance of possible driving mechanisms and present arguments in favor of continent collision as the primary driving force in the development of back arc thrusting behind the eastern Sunda Arc. The azimuth of the slip vector (which is well constrained) for the 1978 earthquake provides strong support for this interpretation. Although the strike of the fault plane for the 1978 earthquake parallels the NW trending bathy- metric expression of the Flores Thrust, the slip vector is oriented nearly north-south (Figure 2), or in the same direction as the relative plate motion between the Indian Ocean-Australia plate and the eastern Sunda Arc (azimuth 0°) [Cardwell et al., 1981].

The distinction between the possible mecha- nisms of back arc thrusting (shallow dipping decollement as opposed to high-angle imbricate thrusting) has important bearing on the evolution of the convergent margin. It is clear from the tectonic development of western Canada, the western United States, and the Andes that both styles of deformation in the back arc region can occur in an "Andean" setting. The implications of either case are even greater, however, when thrusting involves oceanic lithosphere because its negative buoyancy will lead to subduction if a length sufficient to overcome resisting forces is pushed into the asthenosphere [McKenzie, 1977]. Of the two models presented here for the back arc thrust north of Flores, the case of decollement may lead to subduction, whereas high- angle imbricate thrusting would result in crustal thickening possibly followed by uplift and ophi- olite emplacement.

We favor the decollement (subduction) model for the back arc thrust zone. The strongest evidence is the presence of only a single major thrust fault along the length of the thrust zone [Silver et al., 1983]. South of the Flores Thrust there is a noticeable lack of undeformed sediments astride the deformed wedge (Figure 5a) despite the abundance of active volcanic activity to the south. This observation implies that deformation continues within the wedge, but there are no indications of major south dipping faults cutting the deformed wedge, as is predicted by the imbricate thrust model. Thus the deformation within the wedge is probably in response to shear at its base [Davis et al., 1983], along a plane that crops out at the Flores Thrust. The smooth- ness of the gravity field over the deformed wedge also indicates that there are no major crustal offsets south of the frontal thrust.

A second argument is based on the mechanics of thrusting: what is the source of the stress maintaining the gravity lows? The low gravity values cannot be due to thickened crust that remains in isostatic equilibrium and therefore indicates deflection of the underthrusting plate. The load causing deflection will be supported by the plate's flexural strength and therefore associated with a gravity high [Lyon-Caen and Molnar, 1983; Karner and Watts, 1983]. The free-air gravity anomalies over the Flores Basin south of the Flores Thrust become positive at about km 130 (Figure 5a), suggesting that the load deflecting the Flores Basin lithosphere is south of this point. Thus the underthrusting plate (or thrust sheets) must exist beneath Flores Island. The fact that no imbricate thrust

sheets are observed on Flores (faulting beneath Flores Island is predominantly strike slip) is therefore a strong argument against the imbricate thrust model.

Although it is not possible at present to say with much confidence whether or not the observed thrusting is, or will eventually become, a self- sustaining subduction system in the sense that the negative buoyancy of the lithospheric mantle being shoved into the asthenosphere overtakes forces resisting subduction (i.e., buoyancy of the crust, friction on the thrust surface, and resistance to bending), we suggest that the conditions behind the eastern Sunda Arc are favorable for this to occur. Because the arc is still volcanically active, the lithosphere is probably fairly warm and weak, offering dimin- ished frictional resistance to thrusting. In addition, volcanic activity causes high sedimen- tation rates behind the arc, and plate deflection will be infilled by sediments (see Figure 5a) rather than water placing an additional load on the depressed plate. Finally, the northward motion of the Australian continent relative to the Indonesian Islands provides a rapid and long-term driving mechanism for the back arc thrusting. Although little oceanic crust exists north of Flores to feed the newly developing subduction system, further east the Banda Sea contains several hundred kilometers of oceanic crust that could eventually be subducted south- ward beneath the Banda Arc and therefore consti- tute a true arc polarity reversal.

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