Journal of Structural Geology. Vol. 3, No. 2, pp. 143 to 167, 1981
Printed in Great Britain

0191-8141/81/020143-25 $02.00/0

Deformation and metamorphism of the Aileu Formation, north coast, East Timor and its tectonic significance

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(*Received* 21 *March* 1980; *accepted in revised form* 14 *January* 1981)

Abstract—The large block of metamorphic rocks along the north coast of East Timor is of special interest as it occurs at the boundary between continental and oceanic crust in an island arc-continent collision zone. A detailed study of the structure and metamorphic history of 400 km^2 of this formation showed it has a complex history of penetrative deformation but the structure is coherent.

Pelites, psammites and limestones interlayered with dolerites and amphibolites have been metamorphosed in a medium pressure environment. They now form a metamorphic province zoned from low greenschist facies in the southwest to upper amphibolite facies in the east. The earliest recognised deformation phase predated the metamorphism and produced a widespread layer—parallel schistosity but no recognisable folds. The second deformation phase post-dated the metamorphic maximum and micropetrological evidence indicates a gradual cooling during this event. This deformation produced tight folds with an axial plane schistosity and transposed the earlier structures. The progressively weaker third and fourth phases developed crenulation cleavages and related folds, under greenschist facies conditions. Open, fifth phase, macroscopic folds were probably synchronous with strike slip faulting parallel to the north coast. Later dip slip faulting juxtaposed the Aileu Formation with Permian and Mesozoic sediments of very low metamorphic grade.

Reconnaissance K/Ar radiometric dating using hornblende and biotite showed the prograde metamorphic maximum occurred before 11 Ma ago and implies that the second, and strongest, deformation phase occurred in the late Miocene. This young age establishes the relationship of the deformation events to the collision between Australia and the Inner Banda Arc.

The proposed models for the structure of Timor must be modified to fit the deformation history of the Aileu Formation. If Timor is essentially autochthonous, the Aileu Formation was probably deposited in a Palaeozoic graben and the metamorphic maximum may have occurred in the Jurassic. The overthrusting models must be modified in the light of the close correlation in time between penetrative deformation and emplacement of the proposed thrust sheets. The analogy proposed between Timor and 'normal' convergent margins is not supported but it may be possible to draw analogies with the Molucca Sea.

INTRODUCTION

THE ISLAND of Timor, one of the major islands of the Outer Banda Arc, lies within the region affected by the late Cainozoic collision between the northwest margin of the Australian continent with an island arc-subduction complex (e.g. Hamilton 1973, 1977, Katili 1971, Audley-Charles *et al.* 1972, Curray *et al.* 1977, Chamalaun & Grady 1978, Norvick 1979). Discussions of the processes and effects of this collision have so far relied principally on regional geophysics (e.g. Fitch 1972, Denham 1973), marine geophysics (e.g. Beck & Lehner 1974, Curray *et al.* 1977) and reconnaissance stratigraphic-structural appraisals (e.g. Audley-Charles *et al.* 1972, Carter *et al.* 1976, Barber *et al.* 1977, Crostella & Powell 1976, Crostella 1977, Grady & Berry 1977), with only minor palaeomagnetic work (e.g. Chamalaun 1977a, b). Little reference has been made to metamorphic-deformation style in these papers.

Accounts of metamorphism in Timor are derived mainly from a few detailed studies (e.g. Tappenbeck 1940, de Roever 1940, de Waard 1954, 1957, 1959) of apparently allochthonous massifs in West Timor, together with a recent reconnaissance appraisal of the whole island (Barber & Audley-Charles 1976) in the context of plate collision zone tectonics. No detailed studies have previously been reported of metamorphism in East Timor. We are concerned here with examining the metamorphic rocks which occur near the north coast of East Timor and with investigating their possible significance to an understanding of the tectonic development of the Banda Arcs.

The occurrence of metamorphic rocks in this area has been known since the reconnaissance survey of Friedrich Weber, in 1910–11 (Wanner 1956). Of the investigators who later considered this area (Wittouck 1937, van Bemmelen 1949, Gageonnet & Lemoine 1958, Grunau 1953, 1957, Leme 1968, Audley-Charles 1968, Barber & Audley-Charles 1976), most did little more than undertake reconnaissance traverse investigations mainly designed to provide information for regional stratigraphic mapping. Only the work of Barber & Audley-Charles (1976) provided any specific information on the structural-metamorphic history of the Aileu Formation.

In his regional compilation of the geology of East Timor, Audley-Charles (1968) defined the Aileu Formation as a mass of dominantly phyllitic (but schistose to gneissic in the northeast) rocks occupying a large area of northern East Timor, west of Manatuto, and extending up to 30 km south from the coast. Later this definition was amended (Carter *et al.* 1976) to allow for a southward transitional boundary with the previously separately

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defined Maubisse Formation (Audley-Charles 1968). As reported by Barber & Audley-Charles (1976) the Aileu Formation has been subjected to polyphase deformation and to regional metamorphism increasing in intensity to the northeast. A detailed assessment of the geological history of these rocks should provide important evidence in discriminating between rival hypotheses for the tectonic development of the region.

The area chosen for investigation (Fig. 1) extends from the town of Dili in the west, to Manatuto in the east (about 40 km) and lies between the north coast and the Rivers Liho Bani and Laclo, in the south (about 10 km). Thus it encompasses a mountainous ridge running sub-parallel to the coast, varying from 500 m to 1300 m in altitude, and falling away almost to sea level again in the south (except in the southwest corner). Exposures are generally fresh, although discontinuous and restricted mainly to stream beds.

STRATIGRAPHY

Aileu Formation

The Aileu Formation consists of a sequence of pelitic, psammitic, basic and carbonate-rich rocks which have been strongly deformed and metamorphosed. In the southwest, rocks are mainly pelitic phyllite with minor altered dolerite, whereas in the east, amphibolite and calcareous schist are dominant with less common pelitic and psammitic schist (Fig. 2). Because of an absence of clear sedimentary structures, together with the widespread mesoscopic and macroscopic evidence of transposition of layering, the present lithological boundaries, both observed and extrapolated, cannot be confidently interpreted as original bedding surfaces. Thus no order of stratigraphic superposition can be established on the basis of presently available data.

The specific age of the Aileu Formation in the map area remains problematical. Near Aileu (about 10 km south of the map area), Leme (1968) reported Permian fossils in Aileu Formation phyllitic rocks. Further south, from near the previously designated boundary with the Maubisse Formation, Brunnschweiler (1978) reported the presence of Permian ammonites. Furthermore, the Aileu Formation in that particular area is regarded (Leme 1968, Carter *et al.* 1976, Barber *et al.* 1977) as transitional to the Maubisse Formation, which is of well established Permian age (Audley-Charles 1968). However, further west, between Lete Foho and Atsabe, Brunnschweiler (1978) reported the occurrence of Upper Jurassic belemnites in phyllitic rocks near the transition zone. The tentative application of a late Palaeozoic age to rocks of the Aileu Formation on the north coast is supported by the reported occurrence of Palaeozoic-type crinoid ossicles in a marble about 10 km east of Dili (Barber *et al.* 1977). No identifiable fossils were found during the course of this study.

Mapped lithological units

The lithological divisions employed here are of compositional but not necessarily of stratigraphic significance. The boundaries between units are often gradational. "Pelitic phyllites and schists with minor quartzites, quartz schists and amphibolites" is the most common rock type (Fig. 2). It has greater than 70% pelitic material over a minimum distance of 100 m. Almost all rocks south of the Western Laclo Fault (Fig. 2) are in this group. North of the fault there are significant interbeds of quartz-rich rocks. To the east, pelitic rocks still dominate, but a more complex interfingering occurs until in the far east, pelitic rock is uncommon in zones large enough to map.

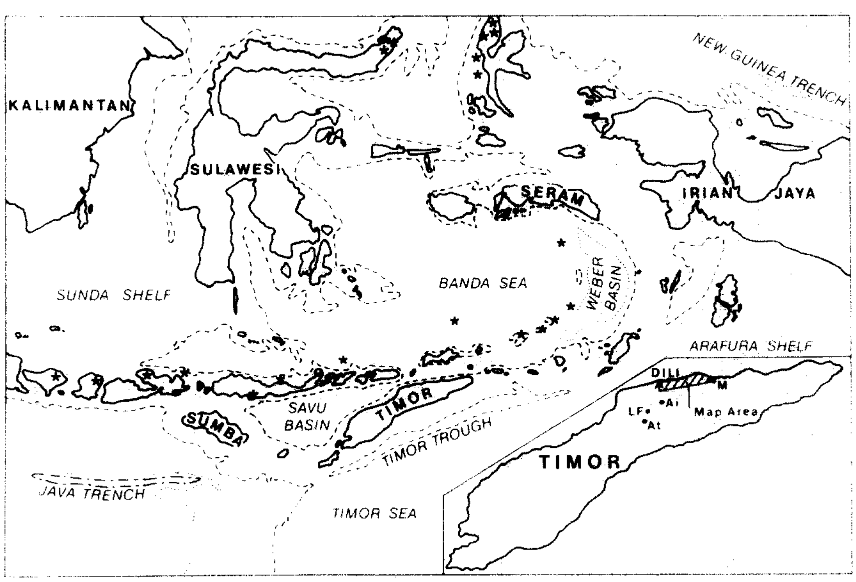


Fig. 1. Sketch map of eastern Indonesia with inset showing location of the mapped area depicted in Fig. 2. Asterisks indicate active volcanoes; even dashed lines, dotted lines and uneven dashed line, indicate the 2000 m, 5000 m and 6000 m bathymetric contours respectively. (Inset: M = Manatuto, Ai = Aileu, At = Atsabe, LF = Lete Foho).

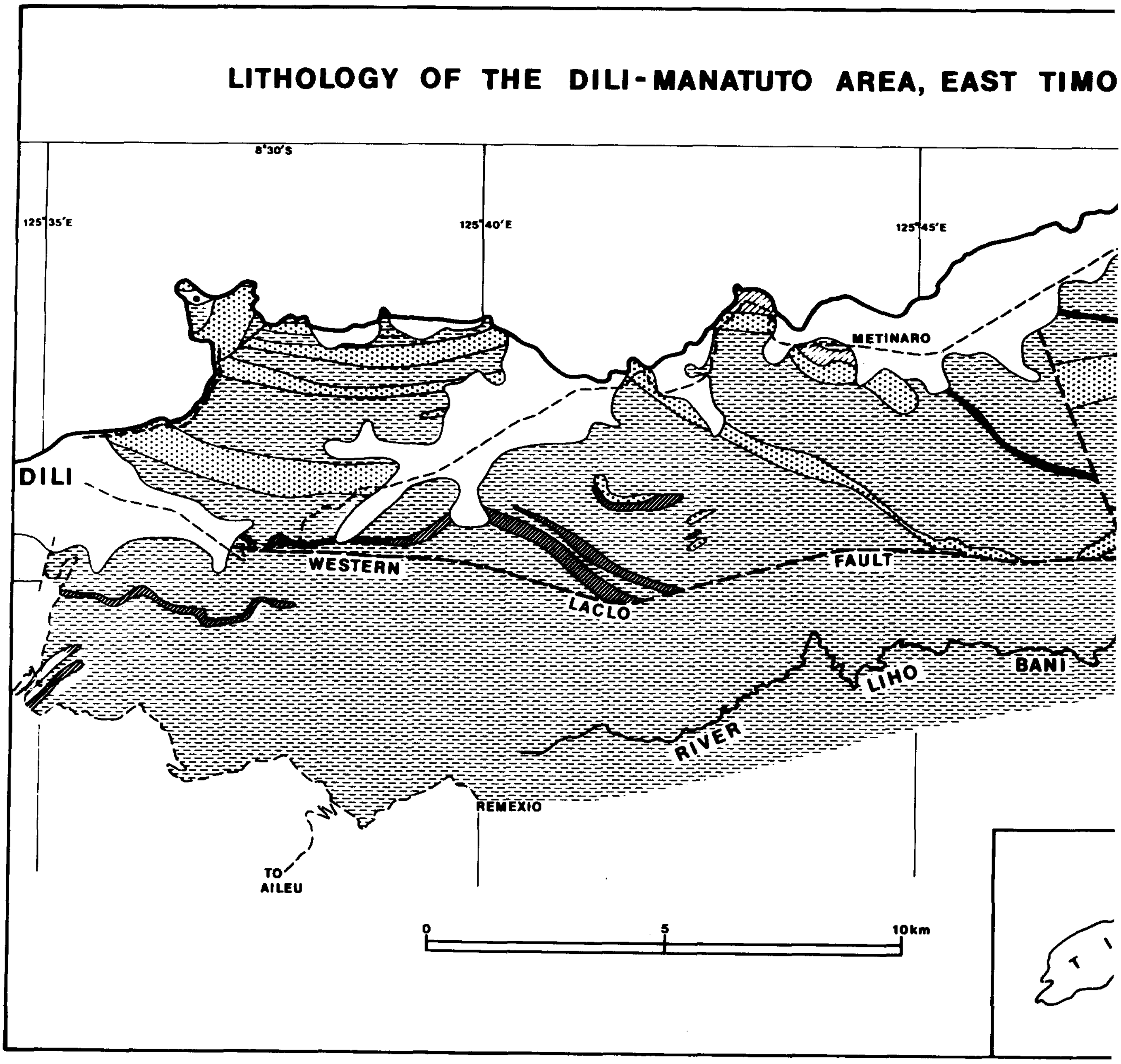
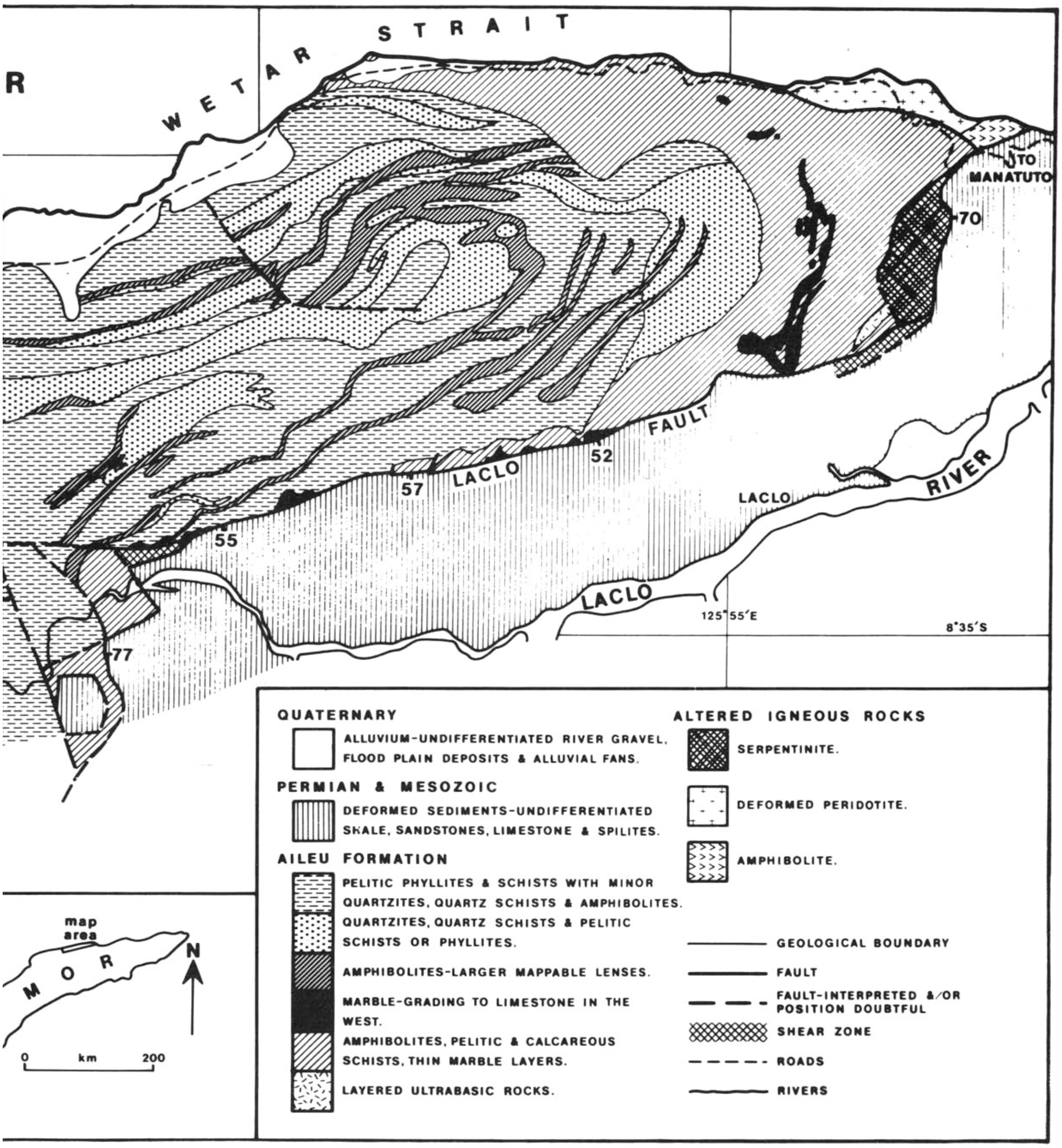


Fig. 2. Geological map of the north coast of East Timor between Dili and Manatut



showing distribution of lithological units in the Aileu Formation.

"Quartzites, quartz schists and pelitic phyllites or schists" includes all sequences with more than 30% psammites, of which quartz schist is the most common. This unit does not occur south of the Western Laclo Fault. In the northwest, it is regularly interlayered with the pelitic phyllites and schists unit with an apparent width of 500 m. The greatest apparent thickness of this unit occurs in the east where it forms a lenticular layer up to 1 km wide.

Amphibolite lenses of more than 50 m apparent thickness are shown in Fig. 2 where they show significant extension along strike. These lenses seldom reach more than 200 m in apparent thickness. Detailed chemical considerations (both major and minor elements) suggest that these rocks are dolerites of alkaline and continental tholeiitic affiliation (Berry 1979).

"Marble and partially recrystallised limestone" of mappable thickness is a minor component in this area of the Aileu Formation. In layers usually less than 100 m thick, it occurs in the east and discontinuously along the Laclo Fault. This lithology is generally coarsely crystalline and white although it grades through pale green and purple to pale green and rarely pink in its western exposures.

"Amphibolites, pelitic and calcareous schists, thin marble layers" occur principally in a large mass near the eastern end of the map area. The major component is amphibolites with a compositional layering about 1 m thick. Thin lenses of pelitic or calcareous schist separate amphibolite layers in many places. Pelitic schists (gneisses in the sillimanite zone) contribute less than 5% of the total rock. Larger lenses of marble have been mapped separately, but most are closely associated with this rock unit. The whole mass is characterised by strong layering and dominantly basic composition.

Layered ultrabasic rocks are exposed at the eastern end of the Aileu Formation outcrop. The layering is an alternation of mafic composition with calcium-rich composition in zones 5–10 cm thick. While the mineralogy is essentially metamorphic (clinopyroxene sphene, epidote, garnet, bytownite, calcite), the layering is parallel to local lithological boundaries and possibly reflects original igneous layering.

The stratigraphic thickness of the Aileu Formation in the map area cannot be accurately determined. The range of metamorphic facies within these rocks (see below) implies that the rocks occupied depth ranging from 10 km (for rock near Remexio) to 20 km (for rocks in the eastern section) at the time of prograde metamorphism. However, a major deformation might have preceded this event, so that estimated depths of burial cannot confidently be used to predict the thickness of overlying sediment.

The earliest work on the stratigraphy of this formation (Wittouck 1937) suggested that a pre-existing block of schists with overlying limestones was intruded by a large diorite body, producing a metamorphic aureole, including alteration of the limestone to marble. This interpretation was not followed by later workers (Audley-Charles 1968, Barber & Audley-Charles 1976), nor does it appear viable from a consideration of the present work. The amphibolites have regional metamorphic textures, and the mineralogy is not compatible with high level contact metamorphism. Marble is an intrinsic part of the sequence and does not sit on top of the metamorphic complex. Furthermore, there is no evidence of post-metamorphic intrusions in the Aileu Formation.

A northern source of sediment for the Aileu Formation has been advocated by Carter *et al.* (1976) and Barber *et al.* (1977) on the basis of a northward increase in psammite component of the Aileu sediments. This study confirms their observations, for the map area, however there is no evidence to suggest that rocks mapped in this area are the specific time equivalent of finer grained lithologies from further south. It is possible that these rocks came from a lower level in the original sedimentary pile. (The northward increase in metamorphic grade gives some support to this possibility).

DEFORMATION HISTORY

The Aileu Formation is strongly and multiply deformed. A model for the deformation history has been erected on the basis of overprinting relationships (everywhere reliable), style characteristics (commonly reliable in this area) and orientation (not commonly reliable in this area) (cf. Hobbs *et al.* 1976, p. 371–375). As many outcrops contain three or more generations of overprinted structural elements, the relative sequence of events is easily determined for the stronger deformation phases. Such information is particularly common in laminated phyllites and schists of the chlorite and biotite zones* (Fig. 3).

Five structural phases are recognised; the three early phases were intense and strongly penetrative on the mesoscopic scale; the late phases were less intense and localised in their development. The characteristics of each phase are outlined below.

Deformation one (D_1); *associated structure*: S_1

The earliest recognisable structure (S_1) is a cleavage, schistosity, or rarely gneissosity, which occurs throughout the area sub-parallel to compositional layering. No indisputable examples of D_1 folds (on any scale) have been observed. Little is known of any possible recrystallisation effects during this phase, because of later pervasive intense metamorphism. The dominant minerals defining S_1 at present vary from chlorite and muscovite in the southwest, through biotite to sillimanite in the east.

S_1 orientation is highly variable in detail due to the very common small scale third phase and to a lesser extent, second phase folds. Its overall dip is steep to the south and southeast (Figs. 4 and 5). Although no macroscopic D_1 folds have been identified, the possibility remains of large isoclinal folds which have not been detected because of

* The mineral zonation referred to here (and later) is that developed during the prograde metamorphic event. It is used in this sense as an identification of the area under discussion and not the metamorphic conditions.

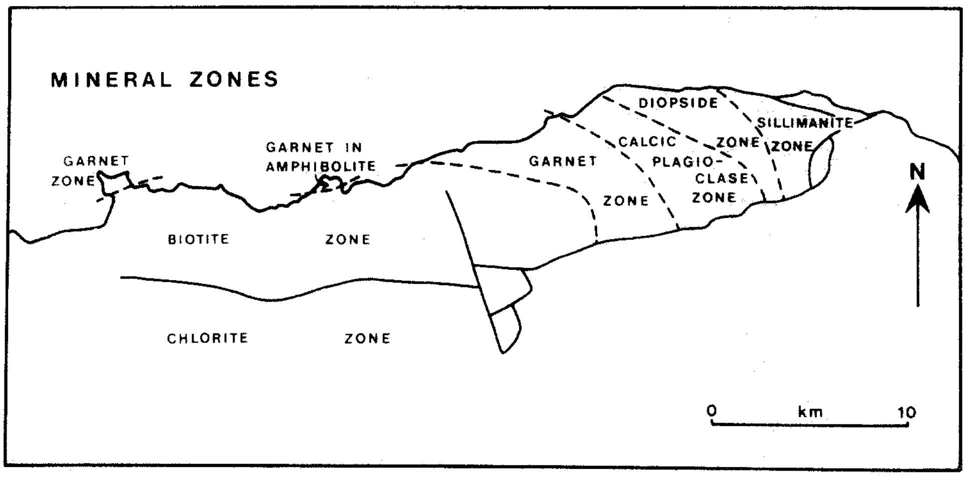


Fig. 3. Mineralogical zones in the Aileu Formation defined from basic and pelitic assemblages in equilibrium during the peak of metamorphism. Biotite, garnet and sillimanite zones are defined by their first appearance in pelitic compositions except for one coastal location where garnet was observed in an amphibolite in an area lacking pelitic rocks. Calcic plagioclase and diopside zones are defined from basic compositions.

the mapping style used (spaced traverses and aerial photo interpretation). Alternatively S_1 may be a bedding plane cleavage enhanced by mimetic crystallisation and D_2 folding.

Deformation two (D_2); *associated structures*: S_2, F_2

The second deformation produced tight and isoclinal folds (F_2) in both lithological layering and S_1. In many places, S_1 is transported into sub-parallelism with the F_2 axial surface, which is also parallel to a new cleavage or schistosity (S_2). Because of these factors, the distinction between S_1 and S_2 is difficult unless both surfaces can be observed in the one outcrop. The minerals defining S_2 are muscovite and chlorite in the southwest and biotite and muscovite further northeast.

Laminated phyllites of the chlorite zone commonly show strongly transposed layering associated with F_2 folds of very short wavelength (usually less than 20 cm) and large amplitude (Fig. 6a). Often limbs are thinned or removed, with quartz-rich material concentrated in hinges (rootless intrafolial folds). These not only form a transposition layering but maintain the linear structure of the original fold axis, forming a rodding lineation parallel to the fold axis.

Finely laminated rocks in the biotite and garnet zones are isoclinally folded without the limb attenuation common in chlorite zone rocks. The wavelength is usually less than 0.5 m, but the amplitude may be more than ten times the wavelength (Fig. 6b). In massive biotite–quartz schists of these medium grade rocks, F_2 folds are close or tight with a new biotite schistosity parallel to the axial surface. These folds usually have a wavelength of 0.5 to 5 m. Folds in the calcic plagioclase and diopside zone have the same style but in the sillimanite zone, a gneissosity is the predominant axial plane feature in some pelitic rocks.

S_2 is subparallel to S_1 (Figs. 4 and 7) and F_2 folds usually drop gently to the south and east (Fig. 8). Lithological mapping and air photograph interpretation suggest a large scale tight F_2 fold closure at Lat. 8°32′S Long. 125°52′E. While F_2 vergence is rarely discernable because of the tight to isoclinal style of F_2 folds, scattered determinations of vergence support the existence of this large closure. No other macroscopic folds have been identified, although for reasons similar to those outlined above for D_1, others could be present in the area.

Deformation three (D_3); *associated structures*: S_3, F_3

Effects of this deformation phase can be observed throughout the map area, principally as mesoscopic folds (F_3), parallel crenulations and often a differentiated crenulation cleavage (S_3), which across the map area generally strikes north–south (Fig. 9). F_3 folds are easily distinguished from earlier folds because of a more open style and contrasting axial surface structure. Both S_1 and S_2 are folded by F_3, but because of similarities in orientation and definition, no attempt has been made to distinguish F_3 folds in S_1 from those in S_2. F_3 fold axes plunge east or southeast throughout the area except in a coastal belt from longitude 125°38′E to 125°46′E (Fig. 10).

The style of folding produced by the third deformation varies systematically across the map area. Laminated phyllites of the chlorite zone have close chevron folds with a differentiated crenulation cleavage parallel to the axial surface(s). In these rocks, the folds, which dominate the outcrops, are quite commonly conjugate or polyclinal (Fig. 6c) and S_3 refracts strongly across lithological boundaries, so that S_3 orientation in individual outcrops fluctuates widely. In the biotite and garnet zones, folds in laminated schists are close and angular (Fig. 6b), but there is less variation in orientation of the axial surface, which can be represented by a differentiated crenulation cleavage. F_3 folds do not dominate the outcrops as they do in the chlorite zone. Within massive lithologies of all three of these zones, F_3 folds are gentle or open in style. Further east, all D_3 mesoscopic structures are poorly developed and the most common expression of this structural phase is a crenulation.

The inhomogeneity and intensity of F_3 folding on all

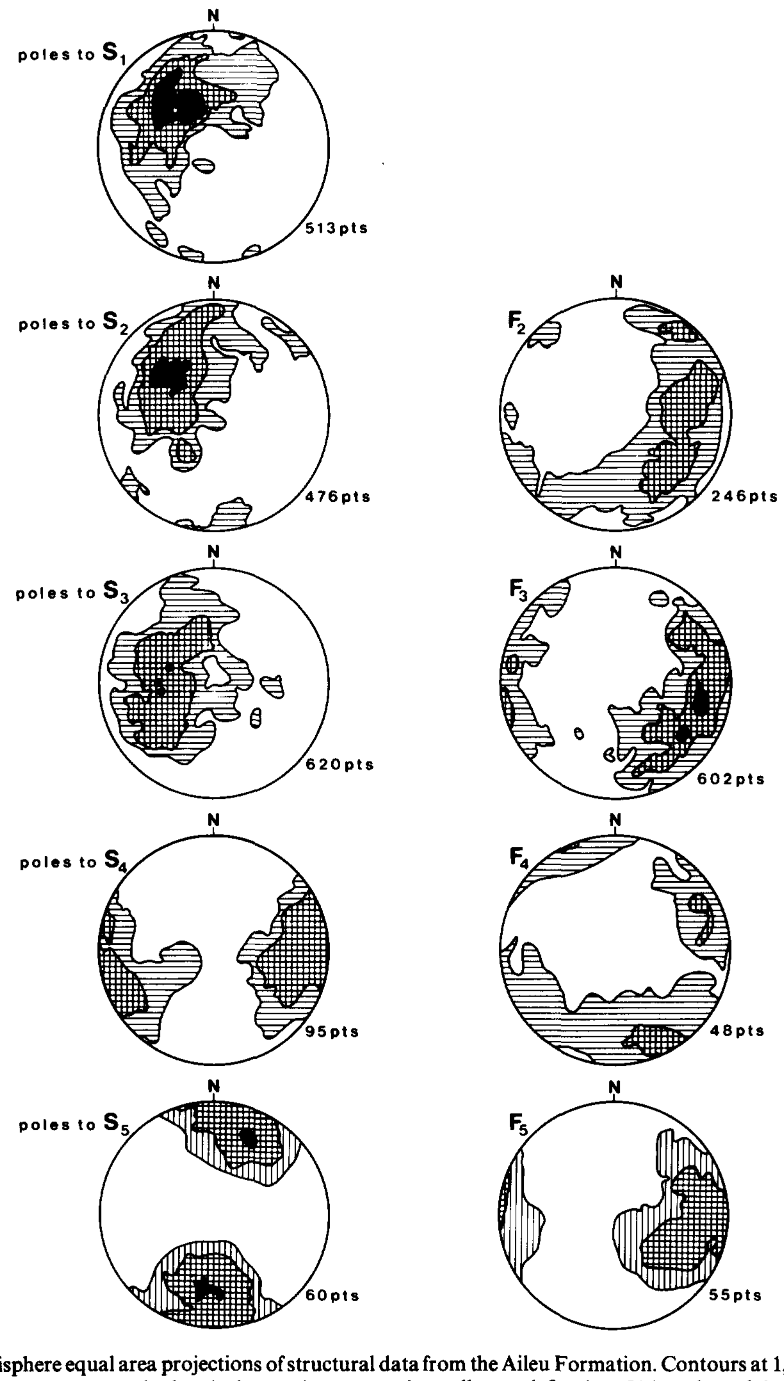


Fig. 4. Lower hemisphere equal area projections of structural data from the Aileu Formation. Contours at 1, 2, 4% per 1% area. Contoured by computer on the hemisphere using a counting cell area defined as 5/ (number of data points +5).

scales has created a complexity of orientation data which is difficult to interpret by traditional methods of domain analysis. To overcome this complication, the orientation data for the first three deformation phases were averaged over map areas 0.5′ latitude by 0.5′ longitude, thus filtering out the small scale variations (Agterburg 1974, pp. 495–497). The results of this analysis (Figs. 5, 7, 8, 9 and 10) while revealing no evidence of significant macroscopic F_3 folds, delineate macroscopic D_4 and D_5 fold closures.

Deformation four (D_4); *associated structures*: S_4, F_4

A set of small, open, angular folds (F_4) of less than 0.5 m wavelength, and crenulations, deform S_3 and earlier surfaces. In many places the conjugate folds of the third phase can be clearly distinguished from these F_4 folds only by overprinting criteria. A set of fine crenulations and monoclinal flexures have the same orientation as F_4 folds, are not folded by F_3 folds and often have an axial plane parting, S_4 (characteristic spacing, 5 cm). Many of these partings contain fine quartz–chlorite–calcite veins.

D_4 structures are most common in the central map area. Over most of the area, S_4 maintains a similar northerly strike to S_3, but has a different dip (Figs. 4 and 11). The orientation distinction between the two phases is most marked on the short limb of a macroscopic F_4 fold in the region surrounding Metinaro. The domain analysis revealed a north-striking macroscopic synform 2 km east of Dili. The corresponding antiform to the east is faulted out (Fig. 12).

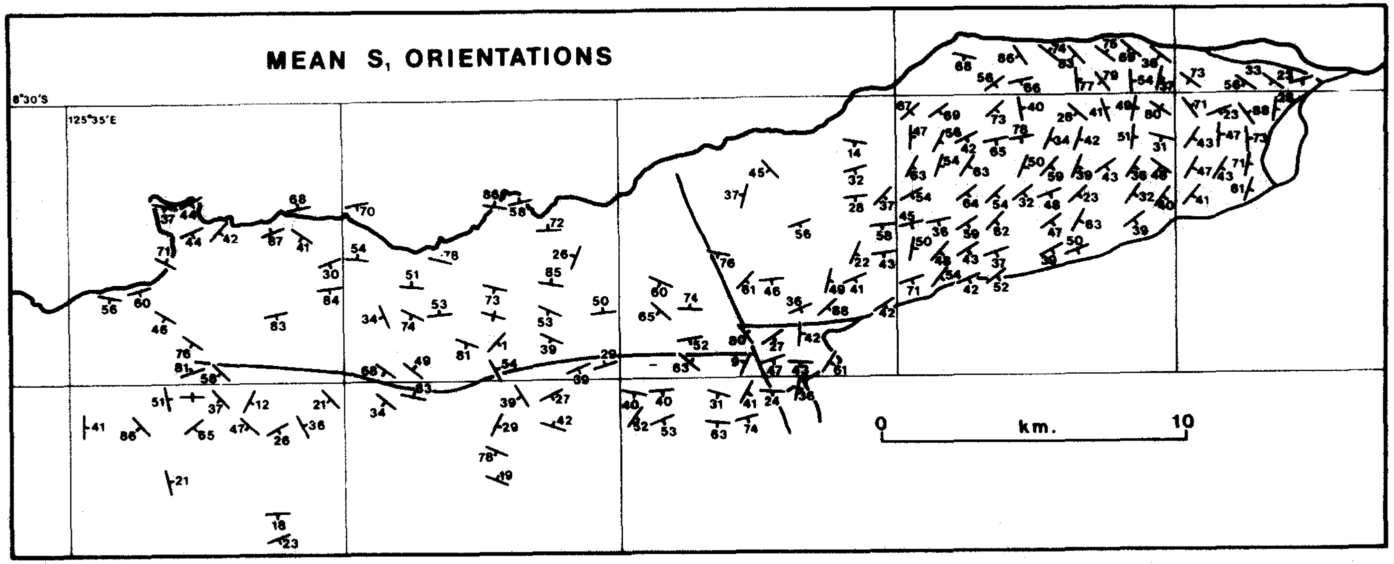


Fig. 5. Orientation of the earliest foliation (S_1) in the Aileu Formation. Each point shown is an average of all the structural data within a square 0.5′ latitude by 0.5′ longitude.

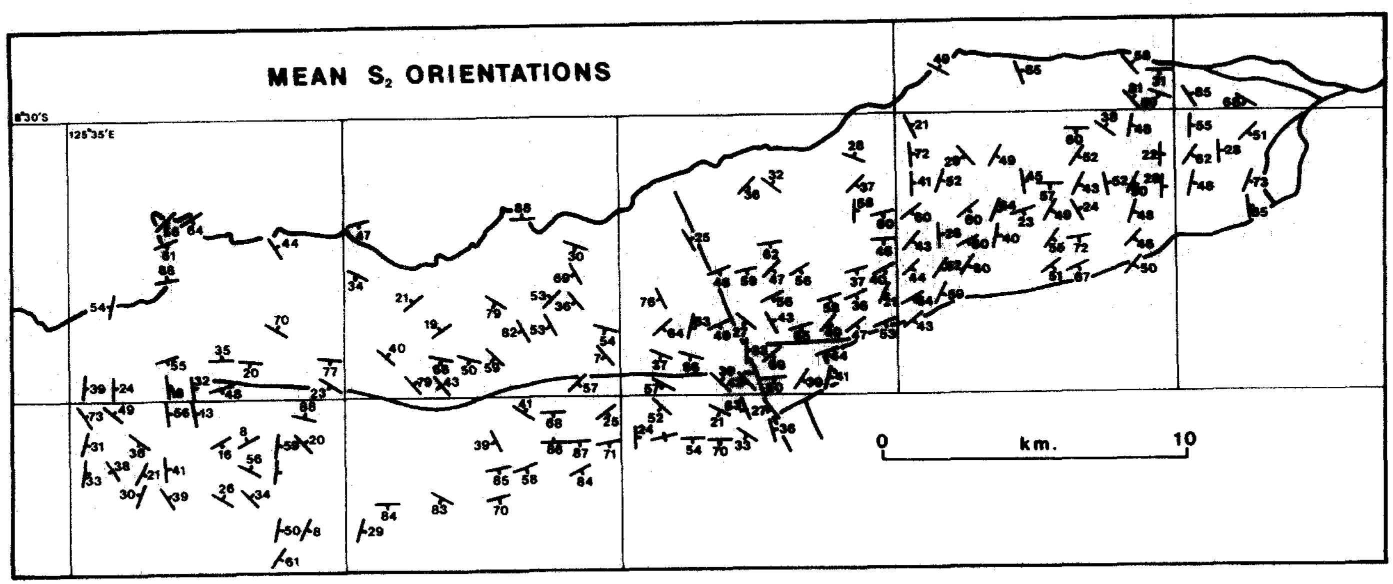


Fig. 7. Orientation of the second foliation (S_2) in the Aileu Formation. Each point shown is an average of all structural data within a square 0.5′ latitude by 0.5′ longitude.

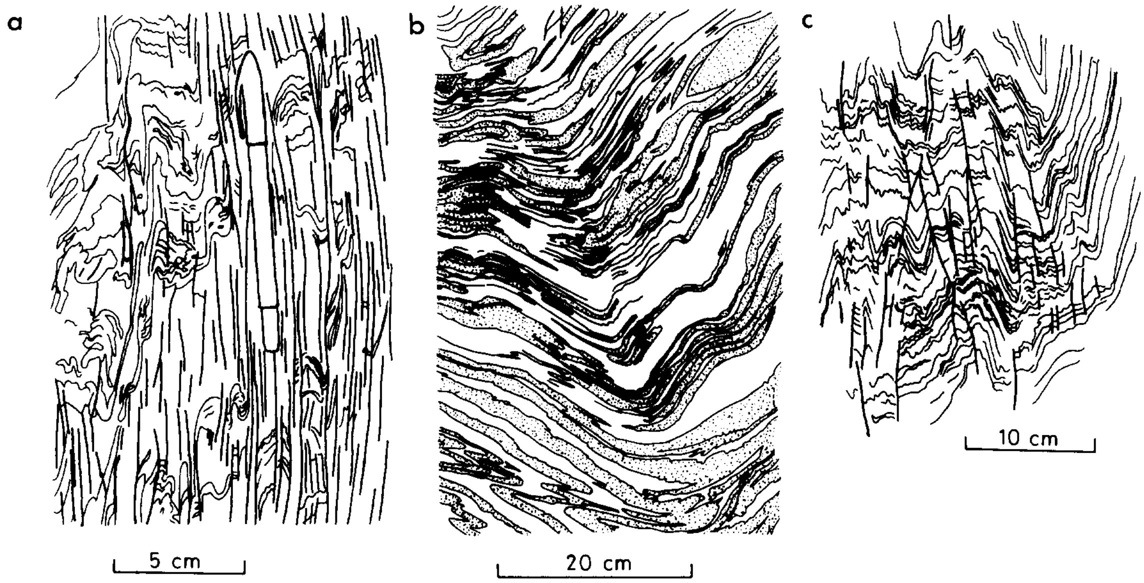


Fig. 6 (a) Transposition of layering in finely laminated chlorite zone phyllites. Differentiated crenulation cleavage (S_2) shown as heavy lines parallel to outline of biro. (b) Tight second phase folds in garnet zone laminated schist refolded by open angular third phase folds. (c) D_3 crenulations in chlorite zone laminated phyllites. Quartz depleted limbs of crenulations shown as heavy lines.

Deformation five (D_5); *associated structures;* S_5, F_5

Fifth deformation folds in the chlorite zone are angular, open to close and of less than 0.5 m wavelength. They rarely dominate the outcrop and the most common expression of this phase is a fine crenulation in S_1 or S_2. In the amphibolite facies region, small folds are rare and the most common D_5 structures are fine crenulations. The axial planes of these crenulations and folds (S_5) generally strike about 100 with a steep dip but show a large spread in orientation (Fig. 13).

Domain analysis reveals several F_5 macroscopic folds (Fig. 12), all with axial surface traces trending east-southeast to southeast. The tightest of these is the macroscopic antiform in the lithological layering, S_1, S_2 and S_3 near the eastern end of the map area.

No direct overprinting evidence exists for the relative ages of D_4 and D_5. F_5 folds overprint F_3 folds and have a distinctive orientation. Indirect evidence, in relation to faulting, helps date F_5 with respect to F_4. A minor fault occurring in the River Liho Bani (Lat. 8°35–26′S; Long. 125°44–77′E) is folded by F_5. This fault is subparallel to the Western Laclo Fault which, in turn displaces a macroscopic F_4 hinge. The Western Laclo Fault appears to also displace F_5 hinges, although the evidence for this is less definite. No examples were found of faults deformed by F_4 folds. These factors, although circumstantial, suggest that D_5 postdated D_4.

Faulting

As with most of the surrounding terrain, the Aileu Formation in the map area is strongly faulted. All boundaries with contiguous pre-Quaternary rock units in the map area are fault contacts. The major faults are shown, with a lettering sequence equivalent to the chronology proposed below, in Fig. 14.

A. Near the road 7 km west of Manatuto, a vertical fault striking 100° forms the northern boundary of the Aileu Formation. The fault plane is clearly exposed in several places along its mapped length. To the north of the fault is a body of deformed peridotite and a smaller body of amphibolite (Fig. 2). Towards the east, the fault and these rock bodies are cut off by the Laclo Fault. Shallow east-plunging slickensides in the exposed fault plane suggest that fault displacement was predominantly strike–slip for part of its movement history.

This fault is probably contemporaneous with the D_5 deformation phase. The fault is parallel to nearby S_5, as defined by the axial surface of the major F_5 fold in the eastern area and this macroscopic fold is the tightest example of a fifth generation fold in the map area. Both the fault and the fold are cut by, and predate, the Laclo Fault.

The present motion of the Australian Plate with respect to the Asian Plate has a left lateral component at the eastern section of the Java Trench (Le Pichon *et al.* 1973). Fitch (1972) argued that decoupling occurs at plate margins, and lateral components in approach direction cause major strike slip faults in the region of convergent plate margins. The present orientation of S_5 is almost normal to the approach direction calculated for the margin by Le Pichon *et al.* (1973). Since the fault is young (i.e. Plio-Pleistocene, see dating discussion below), a comparison with the major strike–slip faults suggested by Fitch (1972) is viable. The movement on this fault may have a left lateral strike–slip component resulting from plate interactions at this convergent margin.

B. Near the eastern end of the map area, metamorphic rocks of the Aileu Formation abut a serpentinite body across a sharp, discordant contact. This poorly exposed fault is probably a steeply dipping, curved (about a steeply

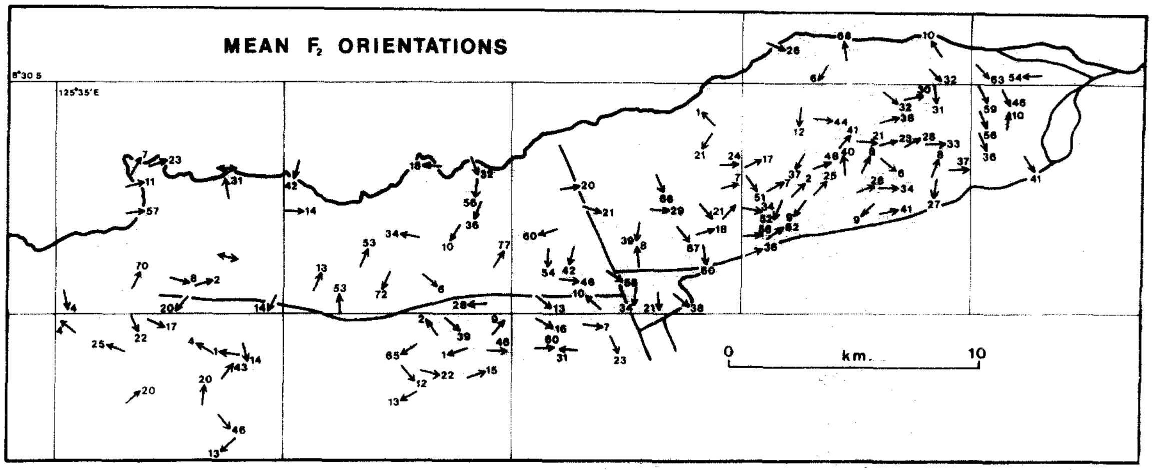


Fig. 8. Orientation of the axes of second phase folds (F_2) in the Aileu Formation. Each point shown is an average of all structural data within a square 0.5′ latitude by 0.5′ longitude.

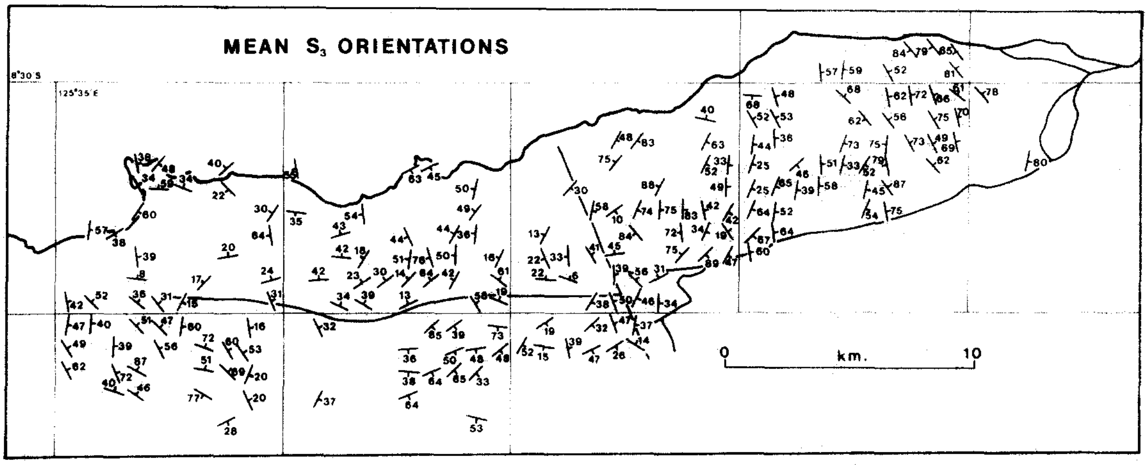


Fig. 9. Orientation of the axial surface of third phase folds (S_3) in the Aileu Formation. Each point shown is an average of all the structural data within a square 0.5′ latitude by 0.5′ longitude.

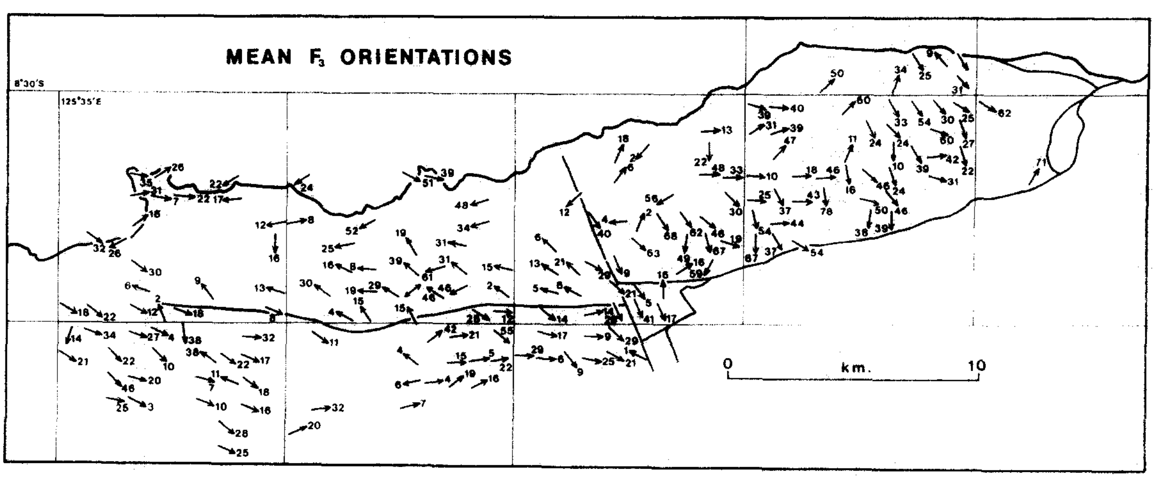


Fig. 10. Orientation of the fold axes of third phase folds (F_3) in the Aileu Formation. Each point shown is an average of all the structural data within a square 0.5′ latitude by 0.5′ longitude.

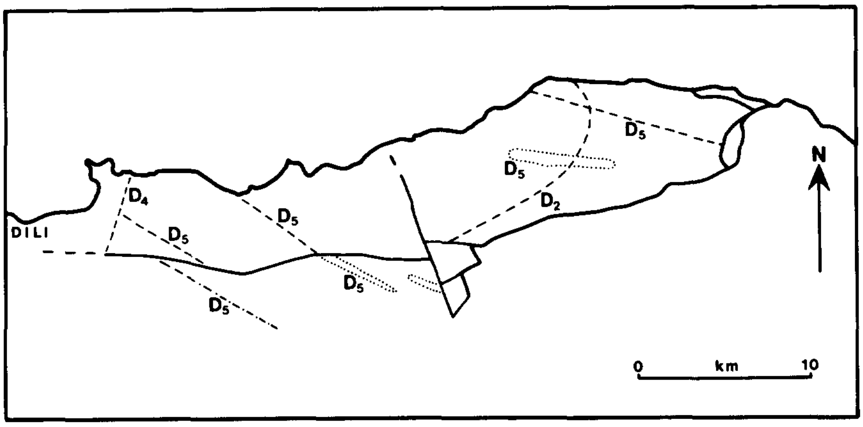


Fig. 12. Location of macroscopic fold closures recognised in this study. Major well established fold closures are shown as dashed lines while smaller and poorly established closures are shown with a dotted line. Each closure is labelled with the deformation phase which produced it.

inclined axis) surface. The fault trace cuts across a strong F_5 hinge (with an opposite sense of curvature) suggesting it is younger than the D_5 deformation. It is cut by, and is therefore older than, the Laclo Fault.

C. The Western Laclo Fault is not well defined by direct field evidence. The principal evidence for it comes from domain analysis, metamorphic mineral zonation and Landsat imagery. The distribution of S_1, S_2, F_2, S_3 and F_3 across the western half of the map area, indicates an east-striking domain boundary north of the River Liho Bani, corresponding with the southern limits of some mapped lithological units. This boundary orientation does not coincide with the orientation of the axial surface trace of any fold which is capable of producing the difference in orientation of structures between these domains. Thus it is probably a fault. The biotite isograd coincides with this domain boundary, thus the fault may be responsible for the rapid W–E change in observed width of the biotite zone. While the line of the Western Laclo Fault does not coincide with any lineament recognised on aerial photographs, the eastern half of the interpreted fault line is represented on Landsat imagery by a distinct lineament.

D. The surface of the Laclo Fault outcrops in many places along its mapped extent. On the easterly-striking section the fault dips moderately (40–60°) south. In the east the fault curves sharply but continuously until it dips steeply (70°) to the east. A small shear zone (containing mixed blocks of serpentinite, metamorphic rocks and Permian–Mesozoic sediments) occurs on the curve of the fault. As the shear zone is small, the motion over the whole fault must have been within the fault's varied orientations. The line common to all orientations plunges 55° towards 145°. At its western end the Laclo Fault is offset by a younger NNW striking fault (type **E**) which is part of a local zone of complex faulting (including another small shear zone).

E. In the central southern area is a group of relatively small faults with complex internal relationships. The data density from this area is insufficient for a complete understanding of the various blocks. Metamorphic grade in these blocks is different from (often higher than) the grade of larger blocks nearby.

All these faults dip steeply and have straight outcrop traces. The small fault-bound block of sediments is not a window through a thrust, as it is enclosed by vertical faults and has outcrops of metamorphic rocks at both higher and lower elevations on all sides.

F. The Laclo Fault is also cut by other minor faults (Fig. 2) which have only small throws, where this can be determined. Further faults of undetermined importance occur within the Aileu Formation. Some can be demonstrated by offsets of mapped lithological boundaries, but others can be inferred only from a consideration of local topographic domains (Berry 1979).

Audley-Charles (1968) determined that East Timor had been affected extensively by block faulting in the Plio-Pleistocene. The Aileu Formation provides very little direct evidence for the age of faulting. One fault-bound, local topographic domain in the map area, is a small plateau (Berry 1979), of the same altitude as the major raised Pleistocene reef complex at Baucau (Audley-Charles, 1968). A type **E** fault forms one boundary of this domain, thus suggesting that type **E** faults were active in the late Pleistocene.

METAMORPHISM

Early geological work on these rocks assigned the strongest metamorphic effects to contact metamorphism (Wittouck 1937). Later reconnaissance geological work showed the metamorphism to be regional dynamothermal, varying from greenschist to amphibolite facies across the area (Barber & Audley-Charles 1976). The results of this study confirm the preliminary conclusions of Barber & Audley-Charles (1976) and, in augmenting them, give considerable new evidence from which a detailed history of varying metamorphism is established, directly correlated with the history of deformation outlined above.

The Aileu Formation has suffered a relatively simple metamorphic history involving one prograde neo-crystallisation event, followed by gradual cooling with retrogressive recrystallisation and neo-crystallisation reg-

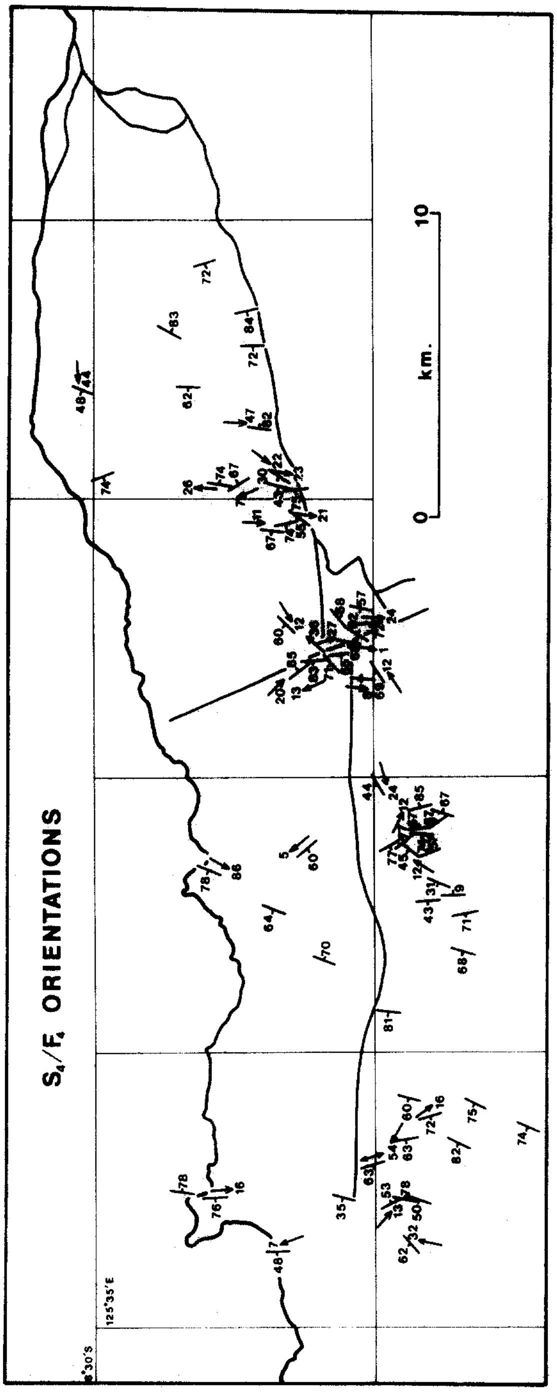


Fig. 11. Orientation of the axial surface (S_4) and fold axes (F_4) of fourth phase folds in the Aileu Formation.

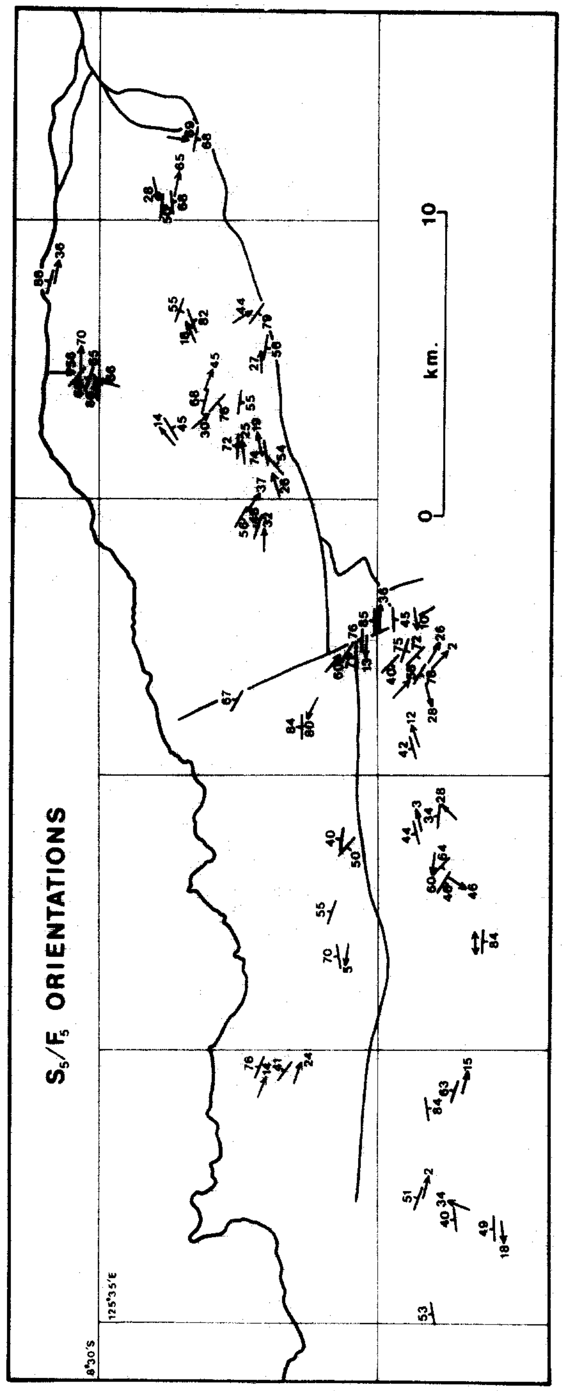


Fig. 13. Orientation of the axial surface (S_5) and fold axes (F_5) of fifth phase folds in the Aileu Formation.

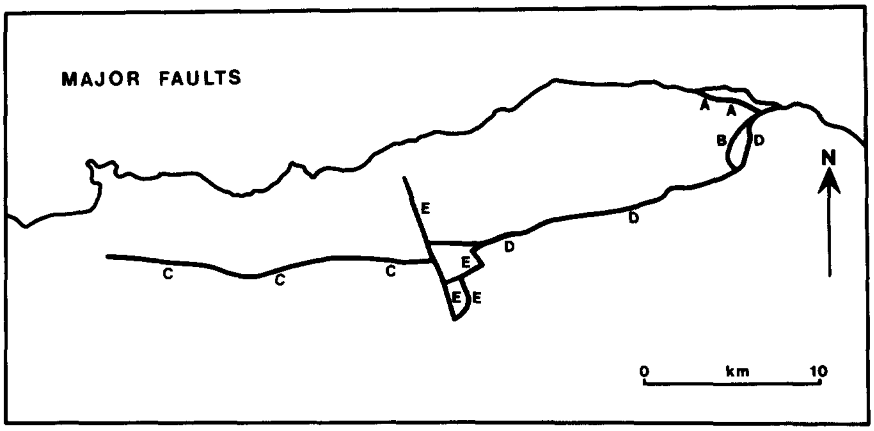


Fig. 14. Major faults in the Aileu Formation. The faults are lettered in chronological order. For further details refer to the text.

istering varying metamorphic conditions. Because metamorphism and deformation are closely related and because evidence for variation of metamorphism with time relates very directly to textural effects of sequential deformation, we will discuss metamorphism of these rocks in relation to deformation phases.

D_1 *to* D_2 *Metamorphism—prograde event*

An intense prograde metamorphic event caused crystallisation and recrystallisation throughout the Aileu Formation. The new minerals and mineral assemblages formed during this phase delineate a metamorphic gradient oriented in general from southwest (low) to northeast (high). This gradient is defined by greenschist facies assemblages (chlorite, biotite and garnet zones) in the south, west and centre of the map area, and by amphibolite facies assemblages (calcic plagioclase*, diopside and sillimanite zones) in the northeast and east of the map area (Fig. 3).

Mineral assemblages developed in the Aileu Formation are:

Greenschist facies

Chlorite zone

pelitic: quartz + albite + muscovite + chlorite ± calcite
quartz + albite + muscovite + chlorite + K-feldspar

basic: albite + actinolite + chlorite + epidote ± quartz ± calcite
albite + chlorite + quartz + calcite

Biotite zone

pelitic: quartz + albite + muscovite + biotite ± chlorite
quartz + albite + muscovite + K-feldspar ± biotite ± chlorite
quartz + albite + muscovite + chlorite

basic: albite + actinolite ± blue green hornblende + chlorite + epidote ± quartz
albite + actinolite ± blue green hornblende + biotite + epidote ± chlorite ± quartz

Garnet zone

pelitic: quartz + albite/calcic plagioclase + muscovite + biotite ± garnet
quartz + albite/calcic plagioclase + muscovite + biotite ± K-feldspar

basic: albite/calcic plagioclase + blue green hornblende ± actinolite + epidote + biotite ± garnet ± calcite ± chlorite ± quartz
albite/calcic plagioclase + blue green hornblende ± actinolite + epidote + sphene + calcite ± chlorite ± quartz

Amphibolite facies

Calcic plagioclase zone

pelitic: quartz + calcic plagioclase/albite + muscovite + biotite ± K-feldspar
quartz + calcic plagioclase/albite + muscovite + biotite + garnet ± staurolite
quartz + muscovite + biotite + garnet + kyanite

basic: calcic plagioclase/albite + blue green hornblende ± quartz ± epidote ± biotite
calcic plagioclase/albite + blue green hornblende + sphene + epidote ± quartz

Diopside zone

pelitic: quartz + calcic plagioclase + muscovite + biotite ± garnet

basic: calcic plagioclase + blue green/green hornblende + diopside + sphene ± epidote ± quartz
calcic plagioclase + blue green/green hornblende + sphene ± biotite ± epidote ± quartz

Sillimanite zone

pelitic: quartz + calcic plagioclase + muscovite + biotite + garnet
quartz + calcic plagioclase + biotite

* "Calcic plagioclase" here implies plagioclase with An > 20.

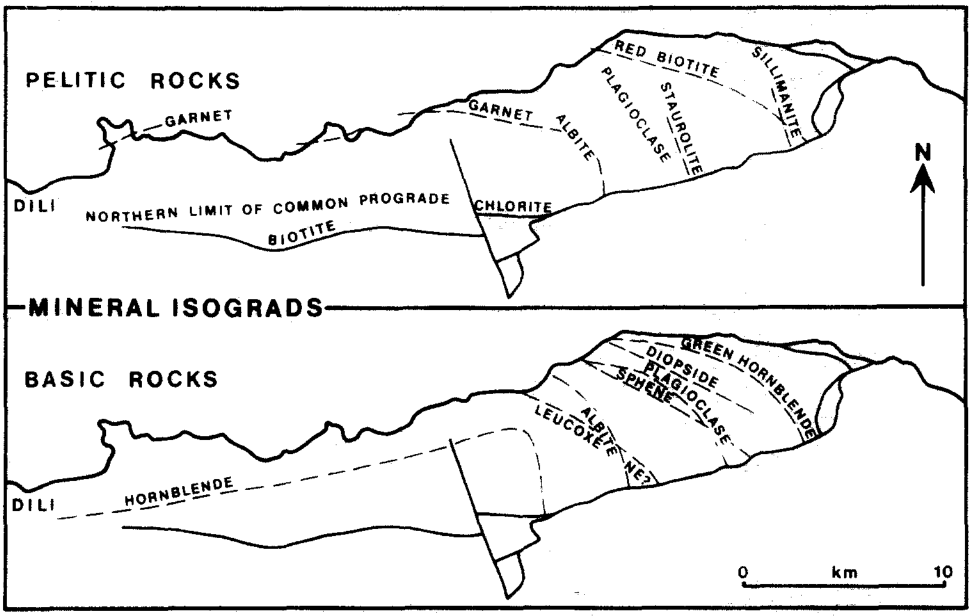


Fig. 15. Mineral isograds defined by assemblages in pelitic and basic compositions from the Aileu Formation.

+ garnet + sillimanite + K-feldspar
± muscovite
basic: calcic plagioclase + green hornblende
+ diopside + sphene ± quartz
calcic plagioclase + green hornblende
+ sphene ± biotite ± cummingtonite

These mineral assemblages imply a wide range of temperatures, but evidence for the pressure at which metamorphism occurred is limited, due mainly to the lack of highly aluminous rocks which normally provide the best control on pressure conditions. Despite these limitations, three lines of evidence suggest that the metamorphic style was of medium pressure type:

(1) In greenchist facies rocks (chlorite and biotite zones) white mica has a b_0 unit cell dimension* of 0.9022 nm, a value very similar to that of the Scottish Dalradian and characteristic of medium pressure type metamorphic provinces (Sassi & Scolari 1974).

(2) The order of appearance of minerals at the greenschist facies–amphibolite facies transition can be used to define the pressure series in a metamorphic terrain. Turner (1968, p. 307) lists the order of appearance of biotite, hornblende, almandine and calcic plagioclase for regional metamorphic provinces covering the known range of pressure. The Dalradian province has a recognisable biotite zone followed by almandine in pelites and by hornblende in basites at the same grade, and finally by calcic plagioclase. At higher pressures, almandine occurs at the same grade or lower than biotite, whereas at lower pressures, almandine and hornblende first occur at the same or higher grade than calcic plagioclase. In the Aileu Formation there is a wide biotite zone and with increasing grade, hornblende appears in basic rocks, followed closely by garnet in pelitic rocks and then by calcic plagioclase (Fig. 15). Thus the mineralogical zonation at the greenschist facies amphibolite facies transition in the Aileu Formation is characteristic of medium pressure (Barrovian) metamorphism.

(3) Amphibolite facies rocks of the map area give few examples of pressure sensitive mineralogy. However, one specimen (Flinders University specimen No. 3-1-158) contains the assemblage muscovite–biotite–quartz–garnet–kyanite. Here kyanite occurs as only trace amounts, but its presence implies that the rock has been metamorphosed at medium or high pressures (Miyashiro 1973, p. 223). The assemblage biotite–garnet–kyanite is used to define the low pressure limit of Bathozone 5 as defined by Carmichael (1978). The upper pressure limit of this zone is defined by the assemblage quartz–Na feldspar—muscovite–sillimanite, an assemblage which is found in some of the pelites from the sillimanite zone. No assemblages characteristic of other bathozones were found in amphibolite facies rocks of the Aileu Formation. Carmichael (1978) states that Barrovian metamorphism lies in Bathozone 5 and suggests that amphibolite facies rocks of this zone crystallised at a pressure of 5–7 kbar (500–700 MPa).

As the rocks of the Aileu Formation crystallised in a medium pressure metamorphic series during this prograde event, the mineral zones are sub-divisions of the "Barrovian-type facies series". They are discussed below with respect to the sub-facies outlined by Winkler (1967).

Chlorite zone rocks have mineralogy typical of the quartz–albite–muscovite–chlorite subfacies. The pelitic rocks contain chlorite, muscovite and minor K-feldspar, indicating a range in composition lying very close to the AFK muscovite–chlorite tie line. The absence of pyrophyllite and chloritoid indicates there are no very aluminous rocks, and the absence of stilpnomelane is probably due to a lack of iron-rich compositions. Basic rocks all plot within the epidote chlorite actinolite field of the ACF diagram.

The first appearance of biotite in pelitic rocks marks the

* b_0 was measured from 060 diffraction peaks, see Sassi & Scolari (1974) (42 determinations, standard deviation 0.0006 nm).

lower boundary of the quartz–albite–epidote–biotite sub-facies. Assemblages are mainly in the field of muscovite—chlorite–biotite on the AKF diagram. The appearance of biotite reduces the compositional field in which microcline is stable, but this effect has not been detected (probably due to limited information on K-feldspar distribution). Mineralogy of basic rocks is unchanged from the previous sub-facies.

The onset of the quartz–albite–epidote–almandine sub-facies is marked by the appearance of hornblende in basic rocks and of almandine in pelitic rocks. These two minerals define isograds that differ slightly, so that the sub-facies shown in Fig. 16 has a lower boundary defined by the garnet isograd, which is more precisely known from the available sampling than is the hornblende isograd. The mineral assemblage found in rocks of this zone have the same compositional range as those of the preceding sub-facies. Basic rocks are still in the same compositional field as the previous sub-facies, except that hornblende replaces actinolite. The low chlorite content indicates the composition lies close to the hornblende–epidote join.

The almandine–amphibolite facies is defined in Winkler's (1967) scheme for Barrovian-type metamorphism by the occurrence of calcic plagioclase. Three subdivisions of this facies are based on assemblages containing staurolite, kyanite and sillimanite. However, staurolite and kyanite have each been found in only a single specimen. Staurolite is stable only in the lowest grade part of the facies (staurolite–almandine sub-facies) and the one occurrence of staurolite lies close to the lower boundary of the facies. The association kyanite–almandine–biotite without staurolite is characteristic of the higher grade kyanite–almandine sub-facies. As no other rocks suitable to distinguish these sub-facies were found, they are combined as a single sub-facies.

The third subdivision is the sillimanite–almandine–orthoclase sub-facies, which is characterised by the assemblage sillimanite–orthoclase–almandine, stabilised by the breakdown of muscovite. As the sillimanite zone in the map area contains pelites with this assemblage, the Aileu Formation has been metamorphosed to the highest part of the Barrovian amphibolite facies, during this event. The wide distribution of sillimanite in these rocks reflects the very wide field of compositions which contain sillimanite, once muscovite is no longer stable; not a different composition of the pelites. In contrast, the rare occurrence of kyanite and staurolite in the previous sub-facies is a result of the small range of compositions in which these phases are stable.

Diopside is a stable phase throughout the almandine–amphibolite facies but is not found in metabasites of the Alieu Formation until significantly east of the lower facies boundary. As garnet is rare in the amphibolites, the rock compositions must lie very close to the anorthite–hornblende tie line. The late appearance of diopside probably reflects a change in the field of stable compositions of hornblende.

The first appearance of green hornblende occurs at the lower boundary of the sillimanite–orthoclase–almandine sub-facies where it is defined by pelitic mineralogy. However, pelitic rocks along the north coast contain quartz–muscovite rather than sillimanite–orthoclase, indicating that the boundary must have a northerly trend. Using the trend of the green hornblende and red biotite isograds (Fig. 15), the lower boundary of this sub-facies has been drawn as far west as sampling allows. The northern extension of the boundary may actually be further east than shown in Fig. 16.

While the surface trace of the mineral isograds and facies boundaries is defined by sampling to ± 1 km the dip of the isograd surface is not known. There is a topographic relief of 1 km in the eastern section, but on the presently available information this can only be used to conclude that if the dip of the isograds in this area is extremely shallow ($< 10°$) it must be a normal sequence and the isograds surface must dip shallowly to the southwest. If the dip is greater than 10° the isograd surfaces must be curved.

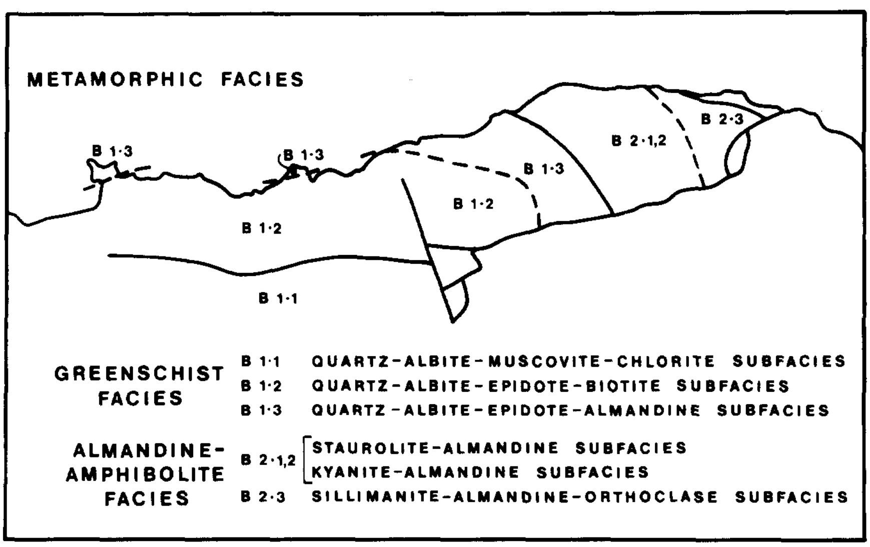


Fig. 16. Metamorphic facies deduced for the peak of the prograde metamorphism in the Aileu Formation. The facies referred to here are from Winkler (1967).

Both S_1 and S_2 cut across the strike of the isograd surfaces. However as S_1 is transposed by D_2 on the mesoscopic scale, it is possible that S_1 was originally parallel to the mineral isograds. If this relationship is assumed, the orientation of F_2 can be used to predict the orientation of the isograd surfaces. In the eastern section of the Aileu Formation F_2 has a moderate easterly plunge. If S_1 was parallel to the isograds they define an antiformal structure with an inverted temperature zonation.

The lithological units mapped in the Aileu Formation (Fig. 2) are much closer in strike to the mineral isograds than either S_1 or S_2. This suggests that the isograds may have been nearly parallel to the primary lithological layering. However the layering has been transposed by D_2 and probably by D_1. Thus the dip of the layering cannot be predicted from the orientation of D_1 and D_2 structures (Hobbs *et al.* 1976, p. 375). The topographic control on the distribution of some of the units, for example the quartzites near the eastern end of the mapped area, provides circumstantial evidence of a shallow westward dip.

AGE OF METAMORPHISM AND DEFORMATION

Microstructural relationships in pelitic rocks across the area allow the relative dating of metamorphism and the deformation phases. The most informative microstructures are associated with biotite porphyroblasts. Three stages can be recognised in the growth of these grains. Biotite porphyroblasts with the basal cleavage parallel to S_1, as defined by muscovite and trains of opaque mineral grains, are heavily included by quartz and opaque minerals (Fig. 17a). They have ragged outlines due to the poikiloblastic growth around contiguous quartz grains. Equant, subidioblastic biotite grains occur which have few quartz inclusions but are still heavily included by opaque minerals (Fig. 17a). In these grains the internal foliation defined by trains of opaque mineral grains is nearly perpendicular to the basal cleavage, and this internal foliation is continuous with the external foliation S_1. The third growth habit is limited to rims on the subidioblastic grains. It is free of all inclusions. Spry (1969) states that inclusion density in porphyroblasts is a function of their rate of growth. Thus the quartz-included biotite grains with the basal cleavage parallel to S_1 probably grew rapidly while the inclusion-free rims on idioblastic grains probably grew very slowly.

The three stages of biotite growth have distinct microstructural relationships. While few rocks not containing a second phase crenulation were found, an important example, sample No. 3-1-162, from near the garnet isograd has both quartz-included and quartz-free biotite porphyroblasts in which the internal foliation is straight, continuous with and parallel to S_1 outside these grains (Fig. 17a). However, as noted above, the two types of porphyroblasts have very different orientations. These microtextures demonstrate post-kinematic, with respect to D_1, biotite growth (cf. Vernon 1978). In this rock the subidioblastic grains have a very thin rim of biotite free of opaque inclusions.

Most pelitic rocks from the biotite, garnet and calcic plagioclase zones contain D_2 crenulations. However the three stages of biotite growth can still be recognised. The subidioblastic porphyroblasts have the basal cleavage subparallel to the axial surface of the D_2 crenulations, and the rims of inclusion-free biotite on these grains show paracrystalline microboudinage (cf. Misch 1969). Where the grains have been strained too rapidly for inclusion-free biotite to grow across the gap there are white mica infillings (Fig. 17b). In these rocks the opaque inclusion trains within the subidioblastic biotite are usually straight. The sum of these textures indicates that the biotite rims free of inclusions grew during the second deformation. There is no evidence to suggest either of the other growth stages continued into the D_2 kinematic stage.

Diopside and sillimanite zone pelitic rocks have different biotite microstructures. For example, sample No. 3-1-78 contains decussate clots of biotite which have D_2 crenulations wrapping around them (Fig. 17c). Some biotite porphyroblasts in other pelitic rocks from the diopside zone (e.g. sample No. 3-1-71) contain curved inclusion trains of opaque minerals which are continuous with a D_2 crenulation outside the grain (Fig. 18a). Such textures suggest pre- and post-kinematic biotite growth with respect to D_2. In addition, very strongly deformed gneissic rocks from the sillimanite zone have finely recrystallised biotite, muscovite and garnet but no chlorite, crystallised during this intense deformation.

Garnet porphyroblasts have a similar range of textures to biotite. In the garnet and calcic plagioclase zone, D_2 crenulations wrap around them (Fig. 17d) indicating they grew before the kinematic phase of D_2. Where the later phase is weaker, S_1 is clearly truncated at the porphyroblast boundary. In the diopside zone, garnet porphyroblasts have been found with curved inclusion trains and in one example (Sample No. 3-1-181) the garnet grains have a strongly curved inclusion train outlining a D_2 crenulation continuous with a crenulation in S_1, outside the grain (Fig. 18b). However, the crenulation in the external foliation is tighter than the crenulation in the internal foliation indicating synkinematic growth of garnet. If the increased tightness is due to a later deformation it may be post-kinematic growth (Vernon 1978).

Sillimanite has crystallised parallel to S_1. The grains have been deformed around D_2 crenulations and show microboudinage features with muscovite infillings (Fig. 18c). Similarly, kyanite is elongate parallel to S_1.

In total the microstructural evidence from the pelitic rocks of the Aileu Formation indicates that the prograde metamorphic event occurred after the first deformation. Biotite crystallised before and during the second deformation in the west and central area, and continued to grow after this deformation in the east. Other minerals confirm this history of the thermal peak occurring before the start of the second deformation and progressive cooling through that deformation. At the end of the D_2 event the temperature in the diopside and sillimanite zones was above the lower stability of biotite and probably below or near the lower limit of garnet stability. In the biotite and garnet zones the temperature had

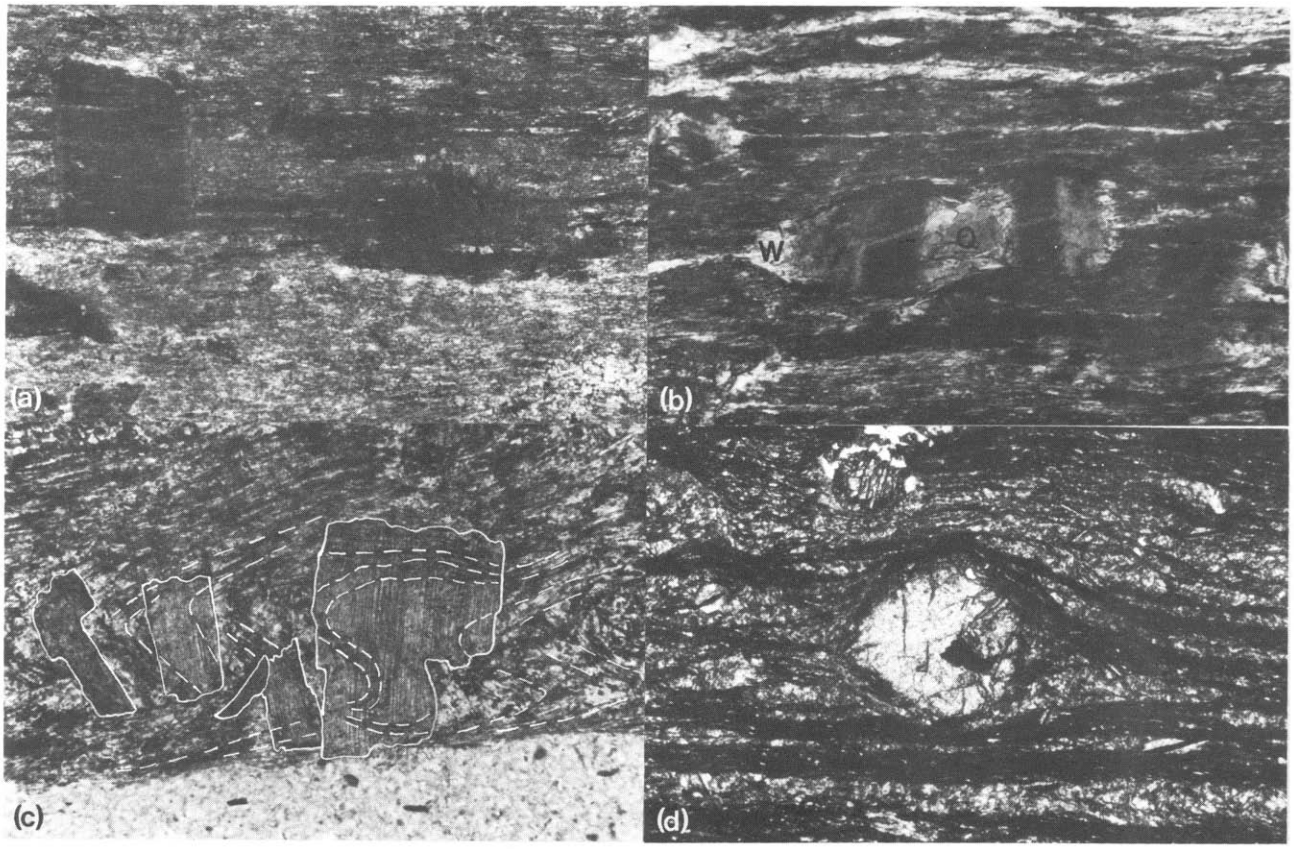
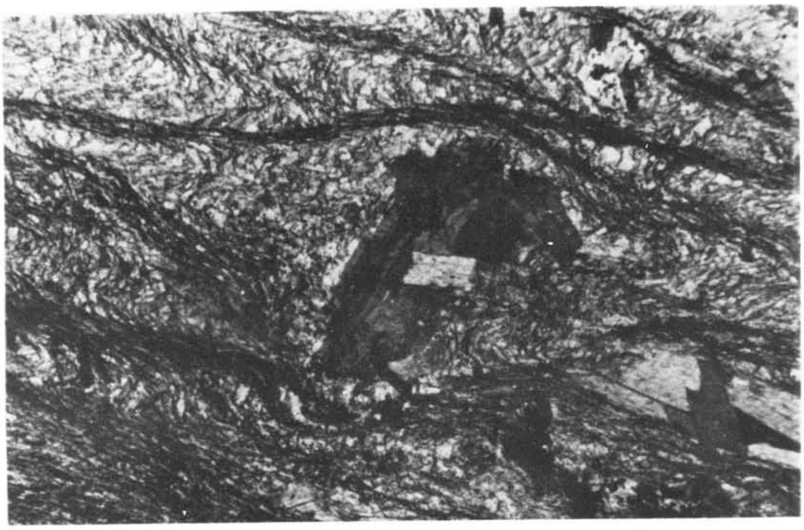


Fig. 17. (a) Biotite porphyroblasts in a biotite quartz muscovite phyllite; as a ragged xenomorphic porphyroblast with the basal cleavage parallel to the cleavage defined by muscovite and lenses of fine grained opaque minerals (at right); and as a subidiomorphic porphyroblast with its basal cleavage at a high angle to the external foliation. In both cases the internal foliation defined by inclusions is parallel to the external foliation. Note the thin inclusion-free rims on the crystal faces of the subidiomorphic biotite (Sample No. 3-1-162). (b) Boudinaged biotite grain. Early crystallised biotite (heavily included by opaque minerals) has been disrupted and overgrowths of inclusion-free biotite have formed. There are also white mica beards (W) which have a cleavage parallel to the external foliation and quartz (Q) biotite and white mica grains have grown as an infilling in the core of the disrupted biotite porphyroblast (Sample No. 3-1-158). (c) Curved internal foliation in the biotite porphyroblasts is continuous with the D_2 crenulation in the external foliation, demonstrating post-kinematic biotite growth (Sample No. 3-1-71). (d) D_2 crenulation cleavage wrapping around a garnet porphyroblast indicating pre-kinematic garnet growth (Sample No. 3-1-141).

a



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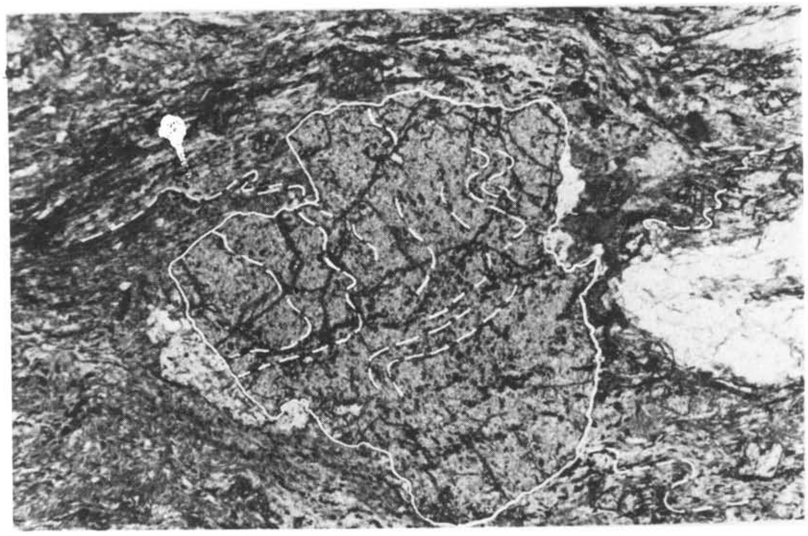

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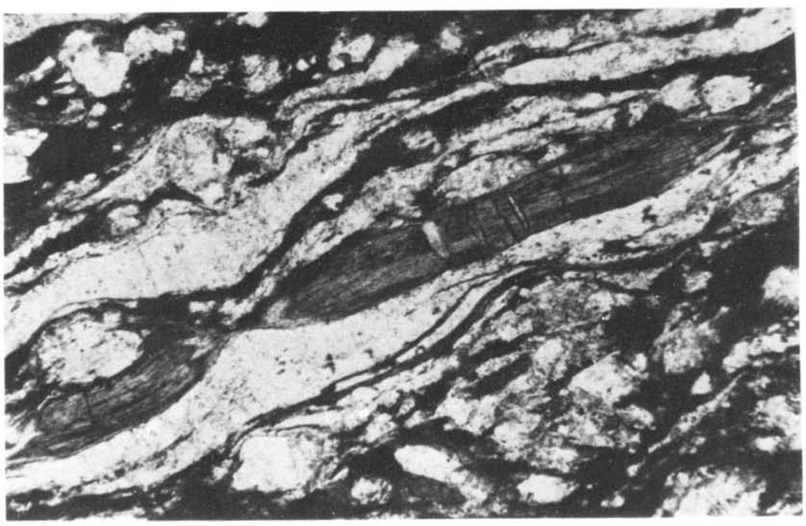

Fig. 18. (a) Second phase crenulations distorted around a decussate biotite clot. The biotite clot predates the development of this crenulation. (Sample No. 3-1-78, plane polarised light, field of view 1.5 × 1.1 mm). (b) A garnet porphyroblast with opaque inclusions defining a crenulated internal foliation continuous with the foliation outside the grain. The second phase crenulations outside the garnet porphyroblast are tighter than crenulations, of the same orientation, in the internal foliation suggesting synkinematic garnet growth (Sample No. 3-1-181), plane polarised light, field of view 5 × 3 mm). (c) Boudinaged sillimanite grain with white mica infillings between microboudins: The sillimanite is heavily altered. (Sample No. 3-1-89, plane polarised light, field of view 1.2 × 0.75 mm).

Table 1. K/Ar radiometric data from the Aileu Formation of East Timor

Sample No.	Mineral	Potassium wt. %	Ar^{40}/K^{40}	Ar^{40} wt. %	Age Ma
1-1-172	Hornblende	0.457 0.454	0.0009682 0.0009565	56.2 51.2	16.6 ± 0.3 16.4 ± 0.3
3-1-186	Hornblende	0.271 0.270	0.00064249	81.8	11.02 ± 0.3
1-2-75	Hornblende	0.384 0.387	0.00063114	73.0	10.83 ± 0.3
3-1-195	Hornblende	0.320 0.322	0.0005993 0.0005939	70.2 67.1	10.3 ± 0.4 10.2 ± 0.4
3-1-192	Hornblende	0.889 0.889	0.00044921	65.7	7.71 ± 0.35
3-1-228	Biotite	7.64	0.00034744	34.8	5.97 ± 0.25

Ages computed using $K^{40}/K^{39} = 0.01167$ atom %, $\lambda_\beta = 4.963 \times 10^{-10}$/yr. $\lambda_e = 0.5811 \times 10^{-10}$/yr. (Analyses by Aust. Mineral Development Laboratories, Adelaide. The analytical procedures and accuracy of this laboratory are discussed in Webb 1976).

dropped below the stability field of biotite.

Microtextural relationships of structural elements produced during later deformation phases (D_3, D_4 and D_5) do not give an exact picture of contemporaneous metamorphic conditions. The third deformation (D_3) produced a crenulation differentiation layering, a common feature of lower greenschist facies pelites (Williams 1972). The association muscovite–chlorite appears to have been stable across the map area. The fourth deformation (D_4) is closely related to quartz, chlorite, calcite veins and zones of retrogressive crystallisation, while no microtextural evidence which limits the conditions during the final deformation (D_5) has been found. Thus, following D_2, the whole area was cooled through greenschist facies conditions, probably reaching sub-greenschist facies temperatures by D_5. The overall evidence suggests a simple single prograde metamorphic event, peaking between the D_1 and D_2 phases, beginning to cool during D_2, continuing to cool during D_3, D_4 and D_5.

Absolute age determinations

There is no conclusive stratigraphic evidence for the age of metamorphism in the Aileu Formation. It is certainly post-Permian and most probably post-Upper Jurassic (because of the biostratigraphic dating on metamorphosed parts of the Formation near Aileu and Lete Foho; Brunnschweiler, 1978). Similarly metamorphism must have occurred before late Pleistocene because of the dating on faults which postdate metamorphism.

In an attempt to provide evidence for the age of metamorphism, a limited programme of K/Ar dating was undertaken on samples selected from this study. The results of six K/Ar determinations, five on hornblende and one on biotite are listed in Table 1 and the sample locations are shown in Fig. 19. Detailed discussion of these results is given in Berry (1979). Five hornblendes from upper greenschist and amphibolite facies amphibolites give ages in the range 7.7 Ma to 16.5 Ma, while biotite from a pelitic schist gives an age of 6.0 Ma.

Three of the hornblende ages are close to 11 Ma (mean 10.7; standard deviation 0.4). Hornblende is very resistant to argon loss and has a high effective closure temperature, close to 480°C, whereas biotite has a lower closure temperature, of about 325°C (Dalrymple & Lanphere 1969, Hanson & Gast 1967, Dallmeyer 1978). One hornblende sample (No. 1.2.75) was collected from close to the garnet isograd, considered to occur at temperatures of 490 to 500°C (Winkler 1976, p. 238). As the hornblende closure temperature is scarcely exceeded by the estimated

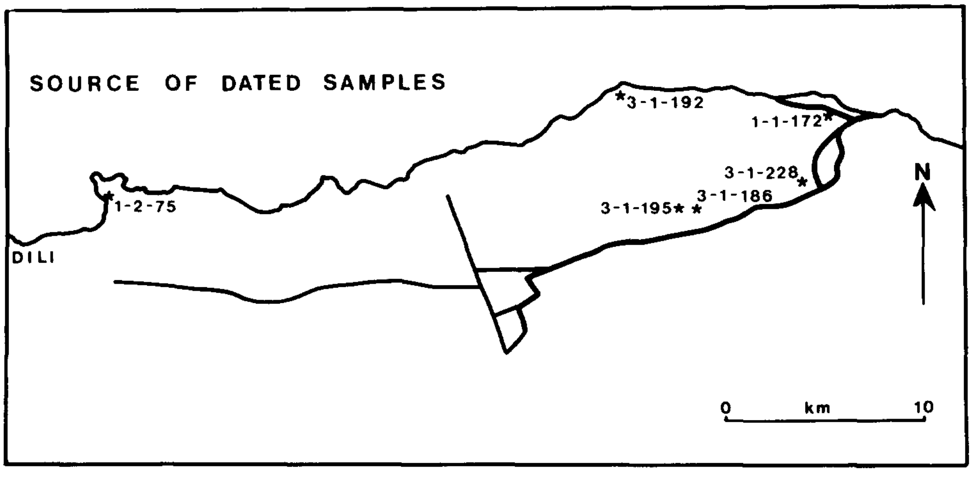

Fig. 19. Sample locations for specimens used in radiometric dating of the metamorphism in the Aileu Formation.

metamorphic peak temperature, the 10.8 Ma date should be a good estimate of the age of that peak metamorphism. Similarly, the two other hornblende samples (No. 3.1.186 and No. 3.1.195) come from near the lower boundary of the amphibolite facies (the former, just above; the latter, just below), indicating a peak metamorphic temperature of about 520°C (Winkler 1976, p. 238). Cooling of only about 40°C is required to begin accumulation of argon in these rocks, thus they are likely to be close estimates of peak metamorphic age. Sample No. 3.1.192, an amphibolite facies rock, gave a date of 7.7 Ma. Hornblende in this rock is a 2 mm poikiloblastic, blue-green variety with a lower 2 V than the hornblende in the other samples dated, suggesting a composition closer to ferropargasite. O'Nions *et al.* (1969), showed that amphiboles near the pargasite end member retain argon at much higher temperatures than ferroparagasites, which are similar to medium grained biotite in this respect. Thus this low hornblende age may be due to compositional variation. The remaining sample gave an age of 16.5 Ma. The significance of this date is not clear, but it may be due to this hornblende gaining a small amount of extraneous radiogenic argon during the cooling history. The date is ignored in the following discussion.

Radiometric dating of biotite in a pelitic schist from the sillimanite zone (No. 3-1-228) indicated accumulation of radiogenic argon began 6 Ma ago. The microstructural evidence from this rock indicated biotite was stable until the end of the kinematic phase of the second deformation. Since the lower limit of biotite stability (400°C, Turner 1968, 350°C, Miyashiro 1973, 420°C, Winkler 1976) exceeds the argon closure temperature for biotite the accumulation of argon could not have commenced until after the end of the D_2 deformation. The second deformation phase finished more than 6 Ma ago.

The microstructural evidence indicates that only one metamorphic event exceeded the biotite isograd temperature, thus the hornblende ages must date the cooling of this phase or an earlier phase. In general, this metamorphism exceeded the closure temperature of hornblende at the positions where samples were taken, so the ages should date the cooling phase of this event. Microstructure also indicate that temperatures cooled to below the biotite isograd temperature by the end of the second deformation. The hornblende ages (except from No. 3.1.192) must relate to the pre- or syn-kinematic stages of the D_2 deformation. While these preliminary results can be considered as tentative, they suggest the prograde metamorphic peak occurred before 11 Ma ago and that the immediately following deformation phase was finished by 6 Ma ago.

Aileu Formation—summary

The Aileu Formation in the study area consists of metamorphosed shales, siltstones, arenites, minor calcareous beds and basaltic-doleritic igneous rocks of alkaline to continental tholeiite type.

The region has undergone a complex deformation sequence, but metamorphism of only one climactic phase. This medium pressure metamorphism ranged from lower greenschist facies to upper amphibolite facies (equivalent to temperatures of 350–650°C, with T/P gradient of 100°C/kbar).

The earliest structural element is a penetrative foliation S_1, everywhere subparallel to lithological layering. Microstructural evidence indicates that biotite porphyroblasts with their 001 cleavage parallel to S_1 postdate it. Metamorphism reached a maximum after development of S_1 and before the start of the second deformation phase. No microstructural evidence exists for high stress in the first phase and no first generation folds, were recognised.

The second deformation caused widespread transposition of layering on all scales. Predominantly pre- (partly syn) kinematic porphyroblasts demonstrate that temperature was dropping during the D_2 kinematic phase. Between 10 and 50°C temperature drop had occurred by 11 Ma, suggesting D_2 started at about the earliest late Miocene. D_2 kinematic phase finished by 6 Ma (late Miocene). The foliation, S_2 produced by the second deformation dips moderately to the south and southeast.

The third deformation produced a foliation, S_3, of markedly different orientation (general northerly strike, steep easterly dip). The indicated contemporaneous environmental temperature was below 400°C. Few structures developed in the eastern section which had previously been metamorphosed to amphibolite facies but the common laminated phyllites in the western section are dominated by close chevron folds (wavelength 5 m), commonly conjugate or polyclinal, and a differentiated crenulation cleavage.

The fourth deformation produced structures of similar orientation to the third, but D_4 mesoscopic structures are not common. Widespread retrogression to chlorite and calcite, and the development of quartz–calcite veins were related to this phase.

Several open macroscopic folds with wavelengths of 5–10 km developed during the fifth deformation phase. Mesoscopic F_5 folds are rare and S_5 strikes approximately 100° and dips very steeply; a major orientation change from S_3 and S_4.

The earliest recognised fault bounds the Aileu Formation on the northeast and might have been active during D_5. Other recognised faults either were initiated, or show evidence of movement, after the entire sequence of penetrative deformation.

TECTONIC IMPLICATIONS

Indonesia has been an important birth place of geotectonic theories during the 20th Century. The island of Timor has been an important element influencing those theories because the outline of the geology has been known and because of the apparent complexity of its geology. That complexity has led to quite intricate models for overall structure and for tectonic development. In recent times a controversy has arisen over elements of the structure and tectonic development of Timor.

Early workers in both West and East Timor generally supported a model involving complex imbrication with widespread allochthonous masses resting on many shallow dipping thrust planes, and overlying autochthonous sequences. Audley-Charles (1968, and in many, more recent papers, with Carter and Barber) maintained the general principles of this model, while augmenting regional structural, stratigraphic and sedimentological detail. During the progression of these publications new details have been added and occasional modifications made, in particular, tectonic interpretations being modified in the light of changes in background geotectonic theory (e.g. the introduction of the plate tectonics–sea floor spreading hypothesis). These workers were prominent in pointing out the close geological ties between Timor and Australia, although consistently maintaining a completely non-Australian, allochthonous origin for material in their thrust sheets.

On the other hand, Hamilton (1973, 1977) and other workers (e.g. Beck & Lehner 1974, Montecchi 1976), were impressed by the early models of complex imbrication in Timor, the apparent continuity between the Sunda and Banda Arcs, and the similarities in sub-bottom seismic reflection profiles from the northern walls of the Java Trench and the Timor Trough. These considerations led them to postulate a tectonic model in which Timor is pervasively imbricated, mixed parap-autochthonous Australian material to allochthonous non-Australian material, occupying the arc-trench gap, above a north-dipping subduction zone which reaches the surface in the Timor Trough.

As a result of multi-disciplinary field observations in East Timor during the early-mid 1970's, another group of workers (from Flinders University) expressed reservations about the complete validity of both major models (expressed most recently in Chamalaun & Grady 1978). As a result, they have postulated an alternative working hypothesis in which almost all of Timor's geology is autochthonous to para-autochthonous Australian continental marginal material, and which is affected only by imbrication with relatively minor overthrusting movement.

These models which we will term the 'overthrust', 'imbricate' and 'autochthon' models respectively, are derived from a complex integration of the geology and in part from analogy with plate tectonic models of convergent margins. The arguments are too wide ranging to treat fully here so we will limit the following discussion to the implications of the deformation history of the Aileu Formation in the context of each of these models.

The 'autochthon model', as advocated by Chamalaun & Grady (1978), requires that all rock material now on Timor, except the Bobonaro Scaly Clay and some other minor units on Timor's extreme north coast, was formed as part of the Australian continent, specifically its northwest shelf and basement. Thus the Aileu Formation is postulated to be composed of shelf, or shelf and slope, sediments which were deeply buried beneath continental margin sediments in the early Miocene and became involved in orogenesis only at the time of the Mio-Pliocene continent–island arc collision. The model further suggests that although some imbrication does exist within Timor, it is a relatively minor feature and more particularly, the movements on individual thrust surfaces or zones were relatively small, perhaps only a few kilometres. The Aileu Formation should not be part of a major thrust slice and should be bounded mainly by steeply dipping faults with large vertical displacements.

Berry & Grady (1981) compared the deformation history of the Aileu Formation with other stratigraphic units on Timor. All pre-middle Miocene formations are multiply deformed as is expected if the Aileu Formation was deformed on Timor. It is possible to see correlates of D_2 and D_3 in these rocks. In addition D_5 probably occurred in the early Pleistocene.

The age of D_1 remains uncertain. It must predate 11 Ma and there is apparently a change in temperature distribution between the time of D_1 and D_2. That is the temperature falls substantially from the prograde metamorphic peak at the eastern end of the area but is very little changed in the chlorite and biotite zones of the western end of the area. This change may be accomplished by a body rotation uplifting the eastern end of the area. However an alternative possibility can be suggested.

Substantial thickness of sediment accumulated on the northwest shelf during the late Palaeozoic and early Mesozoic, especially in graben structures such as the Petrel Sub-basin (Laws & Kraus 1974). Assuming a similar graben had developed in the vicinity of the present north coast of East Timor, the major rifting event in the late Jurassic would cause extensive deformation associated with crustal extension. In addition there would be a metamorphic event associated with the spreading. Since the Aileu Formation was near the new continental margin the geothermal gradient must have been very high in the Cretaceous. Royden *et al.* (1980) provided numerical models for this situation. It is possible using their models to select a degree of 'oceanisation' and rates of post-Jurassic sedimentation such that the temperature of greenschist facies rocks would be relatively constant from the Cretaceous to the Miocene while the upper amphibolite facies rocks cooled by over a hundred degrees during the same period. The sedimentation required is about double the measured thickness on Timor but given structural complications in measuring sections this is not impossible. The arc-continent collision in the late Miocene then effectively ended the metamorphic event by uplifting the Alieu Formation.

This model for the evolution of the Aileu Formation is largely conjectural. However it does explain the observed deformation history. It also explains the difference in metamorphic grade between Permian sediments on the north coast of Timor, sub-greenschist facies, and the Aileu Formation in this region which are upper amphibolite facies. The higher grade areas of the Aileu Formation are interpreted as Carboniferous or even early Palaeozoic sediments over 5 km deeper in the graben than late Permian sediments. The absence of a cleavage parallel to the layering in the Permian rocks may be explained in terms of the response of sediments to crustal extension with increasing pressures and temperature. At high pressures and moderate temperatures there is ductile extension but in the shallow crust the sediments deform by normal faulting (cf. Bott 1973).

The 'overthrust' model, as propounded most recently by Carter *et al.* (1976), Barber & Audley-Charles (1976) and Barber *et al.* (1977), involves the emplacement on Timor of several allochthonous masses, all derived from the north, some being from thousands of kilometres to the north. All recognised metamorphic rocks, the Lolotoi Complex and the Aileu-Maubisse Formations (the latter including some very low grade, to unmetamorphosed rocks), are postulated to have been derived from the Sunda Shelf by rifting and back-arc spreading during the late Cretaceous, forming the proto Banda Sea (Carter *et al.* 1976, fig. 8). These metamorphic rocks became part of the arc-trench gap complex by the early Miocene and were emplaced, together with the other allochthonous masses (presumably along a major sole thrust), over the autochthonous elements of Timor as a result of Pliocene continent-island arc collision.

In this model, the Lolotoi Complex metamorphic rocks are considered to be quite old, possibly Precambrian (Barber & Audley-Charles 1976). In contrast, little information about the metamorphic age of the Aileu Formation is given. By implication the Aileu metamorphism is post-late Jurassic and pre-late Cretaceous (by then it was juxtaposed with now unmetamorphosed Palelo Group rocks, Carter *et al.* 1976, fig. 8).

Presumably the most intense penetrative deformation occurred during this period as the Palelo Formation is not strongly imbricated. This model contrasts strongly with the latest mid-Miocene age obtained from the radiometric dating of the metamorphism and the late Miocene age implied for the strongest deformation phase in the Aileu Formation. A possible alternative is that the difference in deformation between the Palelo and Aileu Formations is a result of their position in the arc trench gap. The deformations D_2–D_5 recognised in the Aileu Formation must have affected the Palelo Formation. Berry & Grady (1981) have argued that the structures reported from the Palelo Formation support this correlation. However the difference in deformation style is not greater than the difference between the Aileu Formation and the allochthonous units. Thus the contrast which has commonly been made between the 'overthrust' units and the Australian correlates by proponents of the 'overthrust' model cannot be supported. Because of the young age of the orogenic events this is not a major problem for the model. The deformations may coincide with and post-date the emplacement of thrust sheets on Timor. Since the preferred age for this event is the Pliocene, that is after D_2–D_4 of the Aileu Formation, some reassessment of the stratigraphic evidence is necessary.

A variation of this model briefly considered by Barber & Audley-Charles (1976) and more comprehensively by Barber (1979), involves the Lolotoi Complex and Aileu-Maubisse Formations being part of an Australian continental fragment that rifted off the northwest Australian continental margin during the Mesozoic development of the Wharton Basin (Larson 1975). It is thought to have become involved with the 'Banda subduction zone' during the Jurassic–Cretaceous, and its subsequent history was the same as that presented above in the main discussion of this model.

A problem with the 'overthrust' model and especially with the variation outlined by Barber (1979) is that a number of tectonic events involved the Aileu Formation well before the Miocene. It is rifted in the Jurassic and then transported along a transform fault during the opening of the Banda Sea. It is then part of the arc trench gap sequence in the early Tertiary. If the Aileu Formation is of Australian origin this model also suggests it was part of a continental fragment accreted to the Sunda Arc in the Cretaceous. All these events only produced D_1. However it must be noted that structural analysis on complexly deformed rocks is an interpretation which outlines the minimum number of deformation events required to obtain the present structure. Evidence of earlier deformations may have been destroyed by D_1 and D_2 events.

The 'imbricate' model propounded, in terms of palaeotectonics, principally by Hamilton (e.g. 1973, 1977) with respect to the development of Timor, has been outlined only in general terms; not being very tightly controlled by stratigraphic–structural relationships published recently for Timor, and relying primarily on accounts of apparent tectonic chaos derived from earlier work in the region.

The main feature of this model is that the whole of Timor forms part of a dynamic arc-trench gap imbricate zone. Involved in this zone is a chaotic mixture of oceanic, continental rise and continental margin material all of which has been dragged down into a subduction zone, has become detached from the underriding lithosphere and by progressive dynamic imbrication has risen up into the imbricate wedge. The underlying north dipping subduction zone is postulated to now reach the surface in the base of the Timor Trough.

Hamilton (1979 p. 127) specifically inferred that the Aileu Formation was metamorphosed by overlying ophiolite suites and has an inverted metamorphic zonation. As discussed earlier this geometry is possible but by no means proved by the data. Hamilton (1979) cites two types of examples for comparison with the Aileu Formation. Woodcock & Robertson (1977) summarised the metamorphic rocks beneath Tethyan Ophiolites. These are medium pressure metamorphic rocks of greenschist and low amphibolite facies. In each case S_1, S_2 and mineral isograds are parallel to each other and the overlying thrust. The metamorphic zones are all less than 500 m, usually less than 100 m, thick and there is a general absence of silicic compositions in the metamorphic aureoles. Williams & Smyth (1973) looked at similar aureoles in Newfoundland. Here the aureoles were all less than 300 m thick. There was a single prograde metamorphic event which continued through D_1 and D_2 but the thermal peak in the amphibolite facies predated D_2. The quartz and feldspar were annealed after D_2.

While the structural petrology of these metamorphic rocks has some similarities with the Aileu Formation, the field relations contrast strongly. The zonation in the Aileu

Formation occurs systematically across 40 km of steep topography. An exceedingly complex and fortuitous distribution of isograd surfaces would be required to limit the true thickness to less than 500 m. S_2 in the Aileu Formation is not parallel to the mineral isograds and there is no evidence that S_1 was parallel to either before transposition. Finally, while the highest grade rocks have a basic composition, siliciclastics are common in much of the Aileu Formation.

Regional metamorphic suites are associated with ophiolites in many areas of the southwest Pacific (Brookfield 1977). These commonly have a complex metamorphic history which is similar to the Aileu Formation. For example the metamorphics of the New Guinea "mobile belt" (Dow 1977) are similar in many respects. However these areas usually include a substantial blueschist terrain immediately under the ophiolite sheet. No blueschist terrain has been recognised in Timor.

The model for an inverted metamorphic zonation in the Aileu Formation is supported by the presence of fault blocks of ultramafic rock near the highest grade areas. However the structure is ambiguous in respect of the isograd orientation and there is not a close similarity with areas where metamorphism occurred because of overthrusting by an ophiolite sheet.

This argument does not prove that the Aileu Formation was not metamorphosed in an arc-trench gap environment. A very similar deformation history has been reported in greenschist facies rocks of Santa Catalina Island (Platt 1976). However the zonation of the Aileu Formation cannot be used as evidence of an overriding ophiolite sheet and the deformation sequence is not characteristic of any particular tectonic environment.

Hamilton (1977, 1979) has developed his model for Timor on the basis of direct correlation with models for arc-trench gap evolution. There are several different models suggested for convergent margins and each postulates a different evolution for the rocks in the arc-trench gap. However all of these have been based on data from subduction zones where the subducting plate is oceanic and has only a thin veneer of sediments. At the same time it is accepted that Timor is the result of the 'subduction' of continental crust.

The processes active in a normal convergent margin may not be relevant when continental crust is being 'subducted'. The structure of the Molucca Sea convergent margins (Silver & Moore 1978) supports this statement. The subducting plate in the Molucca Sea is covered by 20 km of deformed sediment and is thus a close analogy to continental crust. The deformed sediments are not subducted or accreted to the base of the imbricate wedge as predicted by classical subduction zone models (e.g. Karig & Sharman 1975) but are thrust over the imbricate wedge (obducted).

There are two further attractive features of the Molucca Sea margins as an analogy for Timor. Firstly, the axial surface of second phase folds in the Aileu Formation dip to the south which fits the vergence pattern observed in the Molucca Sea but contrasts with the orientation reported from 'normal' arc-trench gaps. Secondly, no high pressure metamorphic rocks have been found on Timor. Blue amphiboles reported in early papers do not imply a high pressure metamorphism as they are Fe-rich (Grady & Berry in prep.). We conclude that the metamorphosed part of the precollision arc-trench gap wedge is not exposed. The Hamilton (1977) and Karig & Sharman (1975) models predict that these rocks should be preserved whereas in a Molucca Sea type collision they are overriden during the collision and are unlikely to be exposed even if they survive later medium pressure metamorphism.

CONCLUSIONS

The Aileu Formation forms a large block of coherently deformed material with no evidence of internal thrusting or imbrication. It was deposited in the late Palaeozoic, possibly in an intracratonic graben, and was metamorphosed at medium pressure with the metamorphic maximum occurring before the late Miocene. A major deformation in the late Miocene produced tight to isoclinal folds with an axial surface dipping to the south. This event marked the onset of retrograde metamorphism. There were three subsequent minor folding events of which the last may be synchronous with strike slip faulting on the north coast. The metamorphic block was juxtaposed with very low metamorphic grade Permian and Mesozoic sediments by dip slip, high-angle faulting during the Pleistocene.

The deformation history of the Aileu Formation demonstrates that arc-continent collision produces a complex multiple deformation history. Most of the rocks on Timor show evidence of a similar deformation sequence to the metamorphic rocks of the north coast. In addition the major deformation in Timor, at least on the north coast, occurred in the late Miocene even though geophysical evidence supports a continuation of the continent-island arc collision to the present day.

None of the proposed models can be discarded on the basis of this new data but each must be modified. The 'autochthon' model has additional constraints placed on it by the early history of the Australian margin. The metamorphic history of the Aileu Formation suggests that the high grade rocks should be pre-Permian and the Aileu Formation probably formed in a major Palaeozoic graben. The 'overthrust' model must be modified to include a major orogeny in the arc-trench gap complex just before the preferred time for thrust emplacement of the Aileu Formation on Timor. It may be necessary to include the Australian margin in this orogenic event. The structure of the Aileu Formation has strengthened the argument against a direct analogy between Timor and 'normal' convergent margins.

Acknowledgements—Our thanks are given to Drs. F. H. Chamalaun and M. J. Abbott for many fruitful discussions on the tectonics of Timor during the project. One of the authors (R.F.B.) was supported during this study by a Commonwealth Postgraduate Research Award and financial support for the field programme was provided by the Flinders University Research Committee. We also thank the reviewers whose comments have helped to balance our discussion.

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