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Alternating thin versus thick-skinned decollements, example in a fast tectonic setting: The Misool–Onin–Kumawa Ridge (West Papua)

F. Sapin [a,c,*], M. Pubellier [a], J.-C. Ringenbach [b], V. Bailly [a]

[a] Laboratoire de Géologie, École Normale Supérieure de Paris, Centre National de la Recherche Scientifique (CNRS-UMR 8538), 24 rue Lhomond, F-75231 Paris, France
[b] Total Head of Structural Geology Department, Total Exploration & Production, CSTJF Avenue Larribau, F-64018 Pau Cedex, France
[c] Total Exploration & Production Borneo B.V., 2nd Floor RBA Plaza, Jalan Sultan, Bandar Seri Begawan BS8811, Negara Brunei Darussalam

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ABSTRACT

The Misool–Onin–Kumawa Ridge (Eastern Indonesia) is a broad anticline in the lower plate of the Seram subduction system. In the south it lies between the Seram accretionary wedge and the young Lengguru fold-and-thrust belt (<8 My). A large seismic dataset from recent petroleum exploration in the area allows the ridge to be interpreted as the result of a dual system of thin-skinned and thick-skinned tectonics. A forebulge effect may be superimposed on the emergent sections of the ridge (Onin and Kumawa Domes), where the morphology has been reactivated.

The evolution results from what appears to be a continuum of deformation through three major stages: (1) formation of a Messinian thin-skinned fold-and-thrust belt over a shaly–silty Permian–Paleocene unit; (2) a Pliocene thick-skinned event responsible for the uplift of the ridge, possibly induced by the onset of continental subduction; and (3) recent Pleistocene deformation when thin-skinned tectonics resumed in the Seram Trough. Currently, the Seram wedge abuts the ridge, transferring compression northward into the Salawati Basin.

The jumps of active detachment levels may be a response to changes in subduction parameters (velocity, rugosity, etc.) during the transition between oceanic and continental subduction, or at least from thinned crust to thicker continental crust.

1. Introduction

The actual structure of mountain belts is greatly dependent on the thickness and rheology of the different layers that compose the crust. Several detachments may be involved and the relative chronology of movement on the slipping surfaces can be interpreted from seismic lines. However, the shift of levels of structural detachment within the same orogen has often proven to be difficult to document.

The structural evolution of a thrust system depends on the stratigraphy, the mechanical properties of the rocks, the duration, the rate of deformation and the uplift versus subsidence ratios (Chester et al., 1991; Fischer and Woodward, 1992; Marshak and Wilkerson, 1992; Doglioni and Prosser, 1997). Therefore, the mechanical properties of the deformed rocks (e.g. competence contrasts) and the parameters of the detachment (dip, depth, friction, etc.) appear to control the final geometry of the structures and the kinematics of the thrust system (Sherwin and Chapple, 1968; Davis et al., 1983; Teixell and Koyi, 2003).

The Misool–Onin–Kumawa Ridge (MOKR) structure displays several detachments and is covered by a fair seismic dataset, with reasonable ties exploration wells. Thanks to field investigations and seismic interpretations, it is possible here to extend surface observations to the upper crust and thus define the relative importance of some of these geomechanical parameters, especially the vertical layering of the crust. We demonstrate in this paper that in this tectonic setting, the concentration of movement on a detachment may be short-lived and shifted quickly vertically from thin-skinned to thick-skinned (shallow sedimentary detachment to crustal (ductile) deeper level). Thin-skinned deformation may resume when the crustal motion is blocked.

Using new field data (Bailly et al., 2008) in addition to seismic interpretation, the objectives of this paper are therefore to: (1) better describe the geometries through cross-sections and a structural map; (2) assess the timing and velocities of the tectonic events which occurred in the area since Mid-Miocene; and (3) propose a model explaining the MOKR formation within simple and consistent regional dynamic settings.

* Corresponding author at: Total Exploration & Production Borneo B.V., 2nd Floor RBA Plaza, Jalan Sultan, Bandar Seri Begawan BS8811, Negara Brunei Darussalam. Tel.: +673 2 229 269x2094; fax: +673 2 237 122.
E-mail address: francois.sapin@total.com (F. Sapin).

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2. Tectonic settings

The Eastern Indonesia region is at the position of convergence between the Eurasia – Pacific/Caroline – Australia plates. It is characterised by high rates of relative movements and oblique motions generally partitioned between subduction zones and large wrench faults (Fig. 1, Puntodewo et al., 1994; Michel et al., 2001; Stevens et al., 2002; Pubellier and Ego, 2002).

In this tectonic setting, high strain rate at the boundaries of fragments is responsible for the formation of structures within a short time span. The MOKR is located between two wedges, the Seram and Lengguru wedges (Fig. 1). The first is a northeastward-verging accretionary wedge related to the subduction of the "Bird's Head" (BH) microplate under the Banda plate. The arcuate Lengguru Belt is a southwestward-verging fold-and-thrust belt (FTB) related to the subduction of BH microplate under the New Guinea block. It is referred to as the "Bird's Neck".

The MOKR is an Early Pliocene NW–SE trending anticlinorium about 700 km long, between Misool Island and the Kumawa Dome. Located north of the Seram Trough, it is a complex structure built during several deformation episodes (Pairault et al., 2003; this study) through the Pliocene. According to Pairault et al. (2003) there was long wavelength folding with uncertain origin in the Early Pliocene upon which was superimposed a forebulge effect due to thrusting of the Seram wedge northwards. However, the wavelength and the height of the MOKR are difficult to reconcile with a conventional lithosphere bending mechanism (Bieniawski, 1967; Turcotte and Schubert, 2002; Watts, 2001; Levitt and Sandwell, 1995). In addition, the succession of events leading to the present morphology of the MOKR, including the proposed folding during the Early Pliocene (Pairault et al., 2003) followed by younger load-induced forebulging needs to be refined.

As shown by Pairault et al. (2003), the non-parallelism between the axial trace of the MOKR (older sediments core) and the axial trace of the Pliocene anticline (MOKR axis) suggests an evolution with at least two stages of deformation.

3. General stratigraphy of Misool–Onin area

The MOKR, mostly a submarine ridge (Fig. 1), is described from oil and gas exploration drillings, 2D seismic lines and outcrops. On of the best examples of the Mesozoic stratigraphy of SE Asia is on Misool Island where it provides an excellent exposure of the rocks. On the two promontories of the western part of New Guinea Island – the Onin and Kumawa Peninsulas – only the New Guinea Lst. and Pliocene clastics are exposed. Its geology is consistent with that of

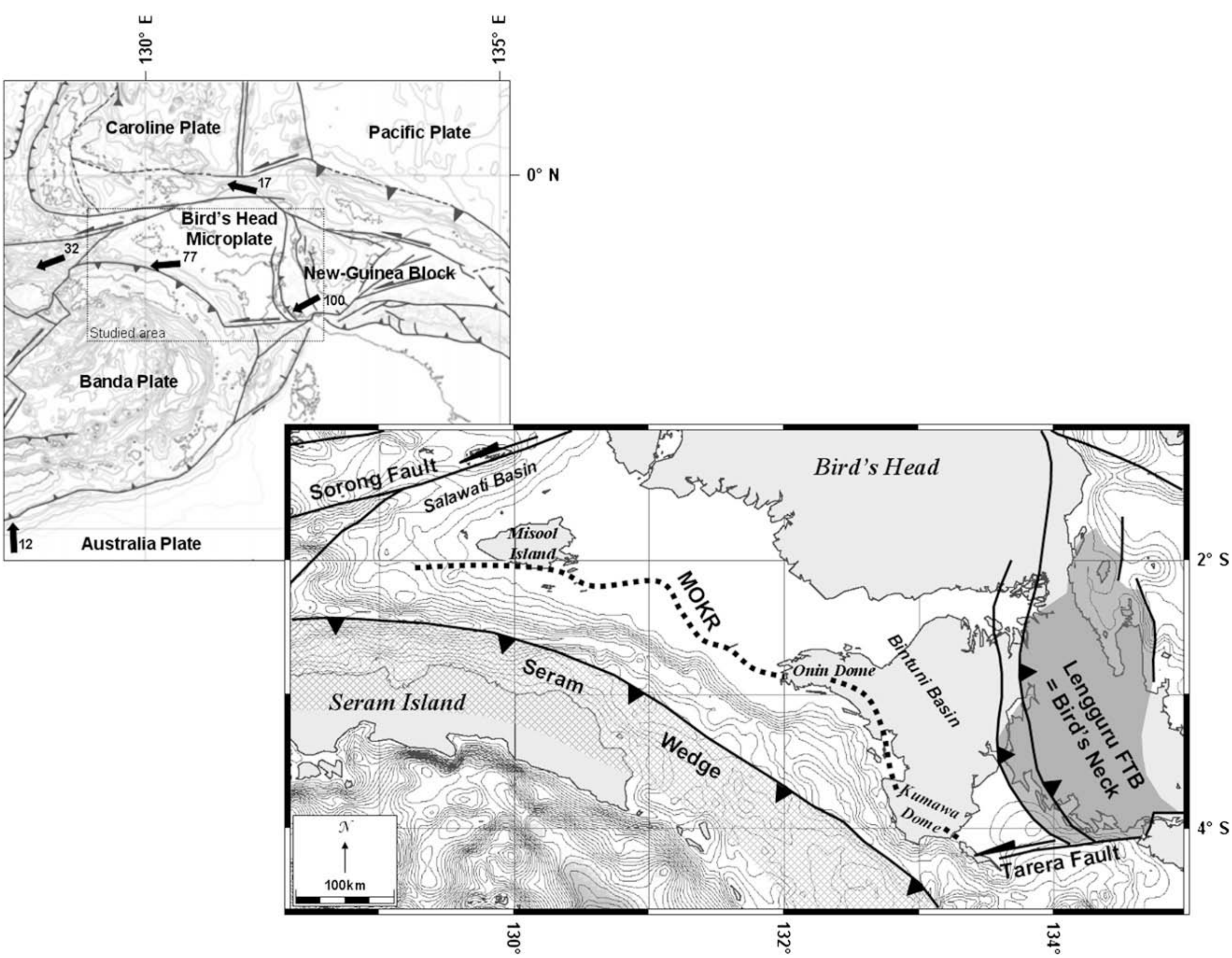

Fig. 1. Tectonic settings of the "Bird's Head" microplate (modified from Pubellier et al., 2000). The velocities (mm/year) refer to the convergence between the crustal blocks surrounding the "Bird's Head" microplate (Pubellier et al., 2000). The boundaries of the "Bird's Head" microplate are: the Sorong fault to the north, the Tarera fault to the southeast, the Lengguru FTB to the east and the Seram wedge to the southeast. A large part (77 mm/year) of the convergence is accommodated by the Seram subduction. The Misool–Onin–Kumawa Ridge (axial trace from Pairault et al., 2003) is a 700 km long structure along the Seram Trough developed on the margin of the "Bird's Head" microplate.

the BH microplate. The stratigraphy (Fig. 2) is similar to the Northwest Australia margin rift-drift sequence and was likely controlled by the break-up of northern Gondwana (Pigram and Panggabean, 1982).

3.1. Stratigraphic units

This stratigraphic summary (Fig. 2) is based on previous outcrop studies on Misool Island and "Bird's Head" (Bailly et al., 2008; Fraser et al., 1993; Pieters et al., 1983; Pigram et al., 1982; Pigram and Panggabean, 1981) correlated with the facies as they appear on seismic lines.

(1) The basement is most likely composed of metamorphosed turbidites as observed on Misool Island (Ligu Fm.) and is inferred to be of Silurian/Devonian age.

(2) The Keskain Fm. (Misool Island) is the equivalent of the Aifam Gp. and the Tipuma Fm. (Bintuni Basin). They correspond to a Permian–Lower Jurassic shaly section. The Triassic Tipuma Fm. is composed of green shales interpreted as syn-rift sediments of the rifting stage of the Northwest Australia margin.

(3) From the Upper Jurassic to the end of the Paleocene, the stratigraphy is commonly described as a succession of dominantly shaly polysequences (Fraser et al., 1993).

(4) From the Eocene to the Middle Miocene, a major carbonate sedimentation episode occurred on the whole margin creating the New Guinea Limestone (NGL). This carbonate episode is divided into two main formations, the Faumai Fm. and the Kais Fm., between which a thin clastic deposit, the Sirga Fm. is locally observed. This has been interpreted to result from the Oligocene New Guinea obduction of ophiolite in New Guinea (Dow et al., 1988; Pigram and Symonds, 1991; Thery et al., 1999).

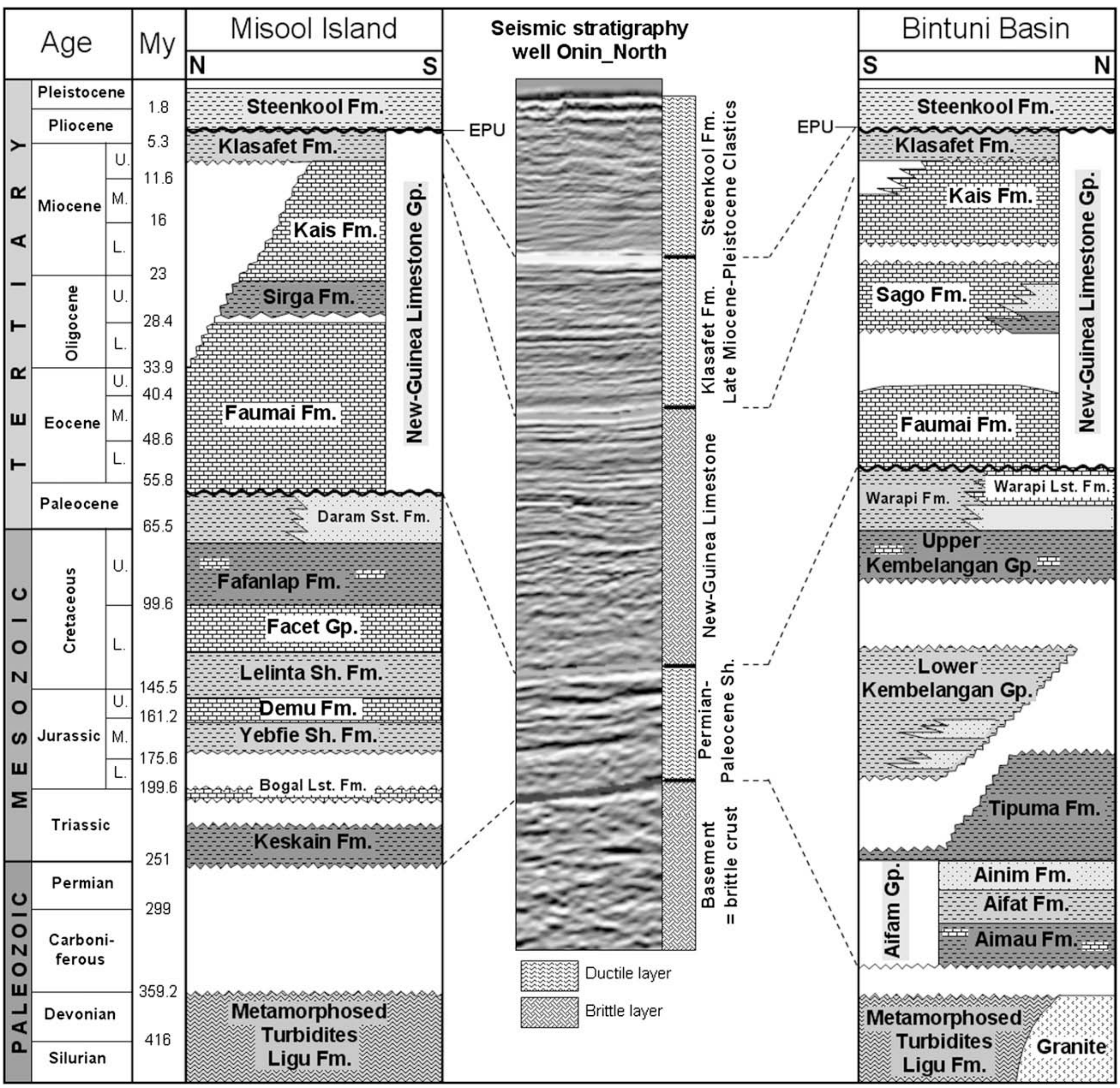

Fig. 2. "Bird's Head" stratigraphy (modified from: Bailly et al., 2008). The well Onin-North is located on Fig. 3. The basement rocks consist of dominantly Proterozoic to Cambrian metasediments and Silurian–Devonian metamorphosed turbidites. The Mesozoic and the beginning of the Tertiary consist of a succession of large dominantly shaly polysequences (Fraser et al., 1993) with highly variable thickness. The Eocene to Mid-Miocene New Guinea Limestone is a very competent layer, nearly 2500 m thick. The Late-Miocene to Pleistocene clastic formations record the tectonic activity in Seram and Lengguru FTB.

(5) Since Late Miocene times, sedimentation changed drastically, with the return of thick clastic formations, mostly silty and shaly (turbidites): the Klasafet Fm. (composing the top of the NGL Gp.) and the Steenkool Fm.

There are three main unconformities (Fig. 2): the base of the Sirga Fm. in Lower Oligocene, related to the Oligocene obduction (see above); the base of the Klasafet Fm. related to the beginning of the Lengguru FTB formation (Dow et al., 1985; Bailly et al., 2008); and the main one, the Early Pliocene Unconformity (EPU), a consequence of the MOKR formation (Pairault et al., 2003).

3.2. Main units and mechanical discontinuities

Because of the implications of variations in the rheology between ductile and competent layers in orogens (Chester et al., 1991; Fischer and Woodward, 1992; Marshak and Wilkerson, 1992; Doglioni and Prosser, 1997; Davis et al., 1983; Teixell and Koyi, 2003), the local stratigraphy has been hereafter divided into three layers based on important contrasts of mechanical properties (Fig. 2). These layers are:

(1) The Permian–Paleocene Shales which are composed mainly of shaly formations. These sediments, which show important thickness variations (from 500 to 1500 m), can be considered, at the scale of the study, as a lower incompetent layer.
(2) The NGL constitutes a very uniform (∼2000–2200 m thick) unit all along the Northwest Australia margin. This competent layer has very high seismic velocities (∼5 km/s), resulting in poor seismic images of the underlying levels.
(3) The Late Miocene–Plio–Pleistocene Clastics, represented by the Klasafet and Steenkool Fms., composed of silty–shaly sediments, represent another incompetent layer above the NGL.

The Permian to Paleocene Shales act as a highly slippery detachment, which allows the propagation of the deformation and controls the superficial structural style.

In addition to the potential detachments within the sedimentary pile, we assume that the crust, as elsewhere, may be separated into two rheological layers, the brittle upper and the ductile lower crust. Hence, the MOKR may include several internal boundaries with significant mechanical contrasts.

4. Morpho-structural characteristics of the MOKR

The MOKR is a broad anticlinorium partly located between two zones of intense shortening (Figs. 1 and 3): in the northeast, the Lengguru FTB; and in the southwest, the Seram wedge. The MOKR

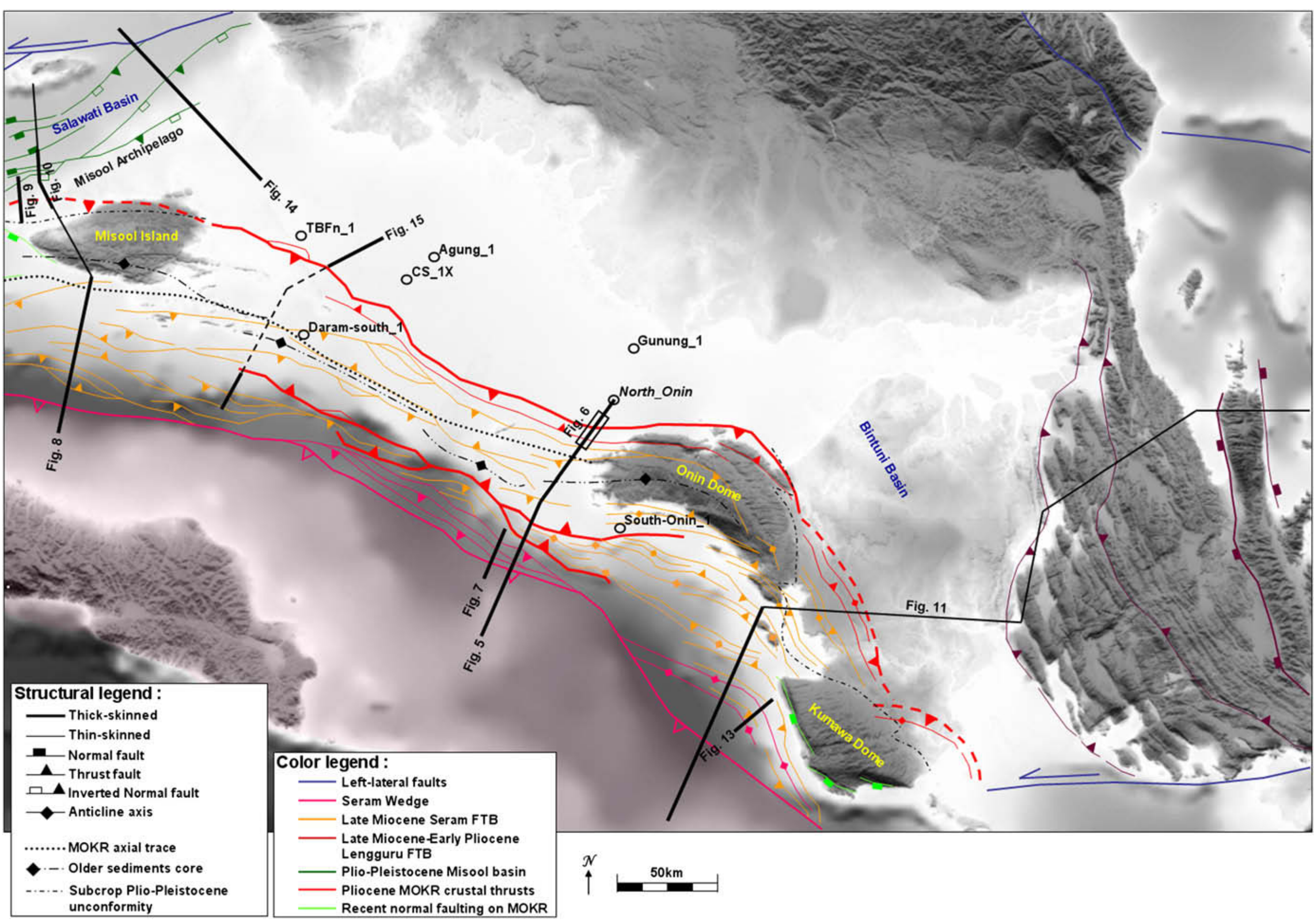

Fig. 3. Structural scheme of the Misool–Onin–Kumawa Ridge on SRTM shaded topography and bathymetry. Location of the wells and seismic lines. The MOKR is a large, 700 km long anticline in front of the Seram wedge. Its northeastern border is prominent thanks to the presence of a southwestward ramp-fold. In contrast, in the central MOKR and Kumawa Dome areas, the southwestern flank is the shorter one. Two epochs of thin-skinned compressive tectonics have been identified: an early one, during the Late Miocene, producing the majority of the folds and thrusts composing the ridge (orange lines); and a recent one where the Seram wedge resumes (southwest Onin Dome and west Kumawa Dome, pink lines). The non-superposition of the MOKR axis (post-EPU) and the oldest sediments core (post-EPU) points to the polyphased formation of the MOKR (Pairault et al., 2003).

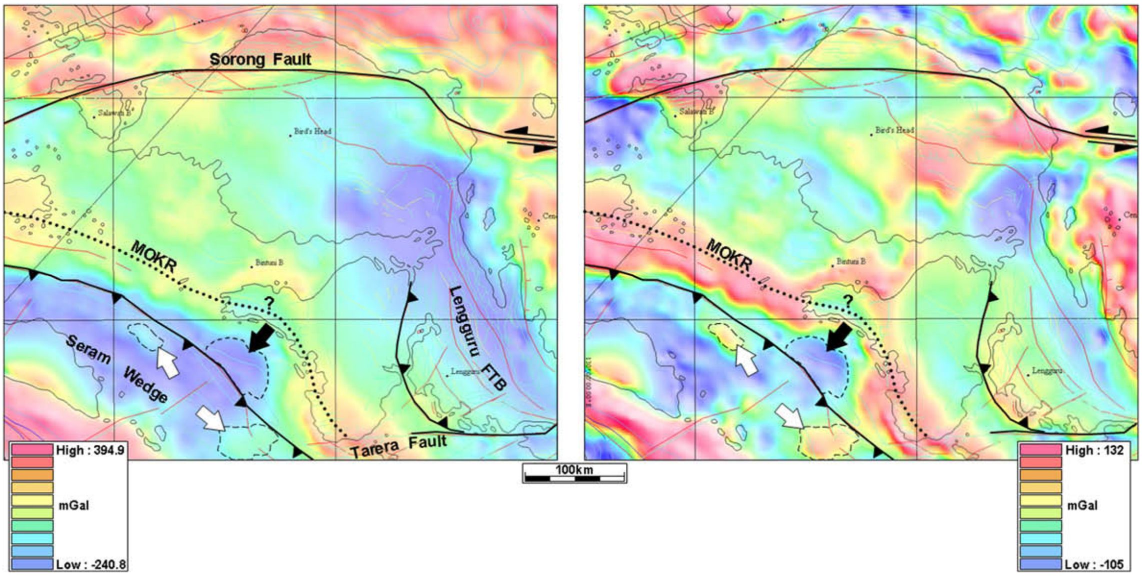

Fig. 4. Bouguer anomalies (left) and 250 km high-pass filter (right) in "Bird's Head" area. The positive Bouguer anomalies are highest in the south (Kumawa Dome) possibly because of the young topography. The MOKR is well-marked on the high-pass filtered Bouguer anomalies; underlining the fact that the ridge is likely to be rooted on deeper levels of higher density, in the crust (or lithosphere). They face – in the southwest and the east – the large negative anomalies of the Seram and Lengguru wedges. The dashed curves signal the presence of abnormally high anomalies within the Seram wedge (white arrows). They are more prominent on the high pass filtered Bouguer anomalies and may be interpreted as subducted asperities. The negative anomaly in the south of the Onin Dome (black arrow) is a consequence of the presence of non-bulged part of the Seram FTB above (Fig. 3).

is a 700 km long structure, oriented NW–SE, and composed of two types of folds with two distinct wavelengths (Pairault et al., 2003; this study; Fig. 3): (1) a series of pre-Pliocene thrust-folds of small wavelength (this study); and (2) a broad anticline, affecting the folded layers and the lower shallow units (Pairault et al., 2003) of the Steenkool Fm.

The entire area may be separated into three segments having different structural trends and morphologies. Despite these

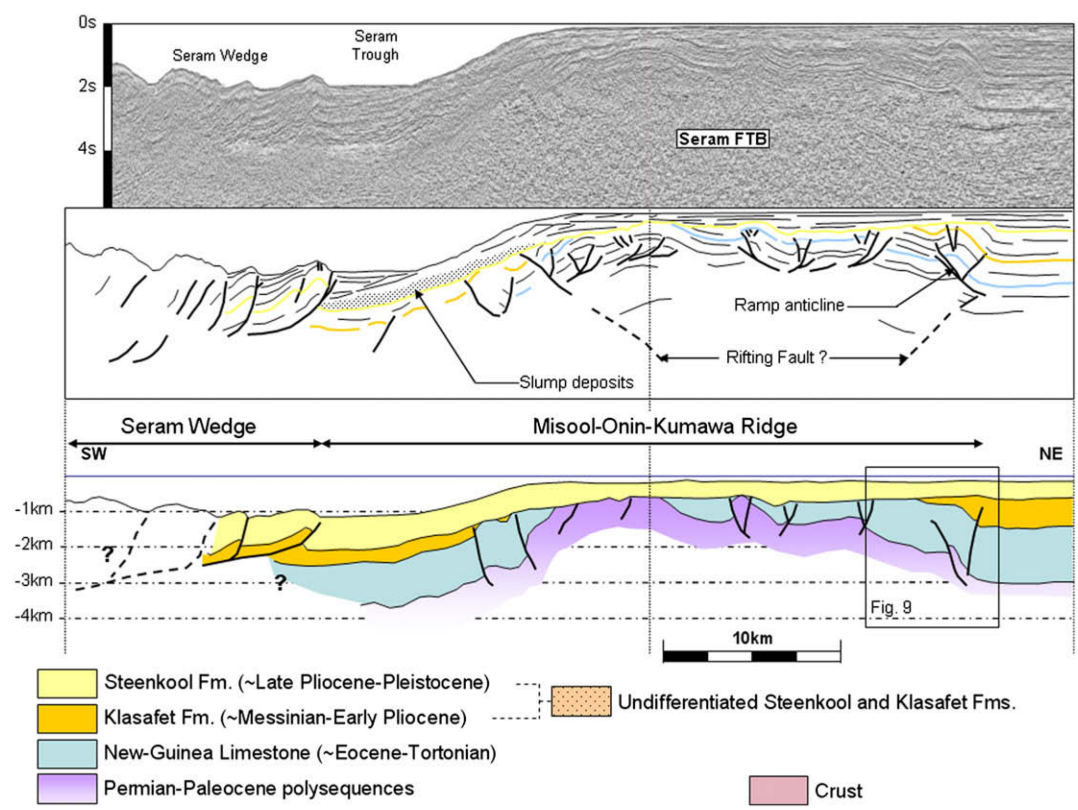

Fig. 5. Cross-section 2: seismic interpretation (top) and cross-section of the MOKR inverted short flank in the central zone (bottom, vertical exaggeration x2, location on Fig. 3). In the central MOKR area, the southwestern flank is steep. In the Seram Trough, the faults of the wedge affect the youngest sediments.

morpho-structural differences, these areas are associated with a single high Bouguer anomaly (Fig. 4) and share a common characteristic southwestward ramp-fold marking the northeastern border of the MOKR (Fig. 3). The presence of a high Bouguer anomaly (Fig. 4) beneath the MOKR suggests either this large scale structure has not yet reached isostatic equilibrium or it is underlain by higher density crustal material.

The short wavelength thrust-sheets are between 5 and 10 km width (Fig. 3) and affect the lower sedimentary units: the Permian–Paleocene Shales and NGL. The detachment level is within the Permian–Paleocene Shales and although the main vergence is to the northeast, many back-thrusts are observed. Although they affect the entire MOKR, they are easily observable only where the Klasafet Fm. is well-preserved, for example in the south of Onin Dome (this study). The negative Bouguer anomaly in the south Onin Dome area (Fig. 4) is likely to be a consequence of these folds.

The MOKR has a width varying between 50 and 150 km. This is a large wavelength structure, compared to the kilometre-scale folds of the sedimentary wedges which deform the EPU. The frontal structure that marks the anticline is a southwestward-vergent ramp-fold (Fig. 3) above a normal fault (this study). This structure can be traced from Onin Dome where it crops out to the east of Misool Island.

4.1. Central area – high dip of the southwestern flank of the MOKR to the Seram Trough

In the central area, the MOKR is only 50 km wide (Fig. 3). The southwestern flank dips strongly towards the Seram Trough ($\sim 10^{\circ}$). It becomes the short flank (Fig. 5) whereas the northeastern flank is characterised by a kink fold (Fig. 6), the only fold not sealed by the EPU.

The Seram wedge (Fig. 7) is associated with a negative Bouguer anomaly (Fig. 4) – due to the thickness of sediments of lower density ($\sim$2.1 sg) than the average crust (2.7 sg). However, to the south–southwest of the MOKR central area, we observe a high Bouguer anomaly (Fig. 4).

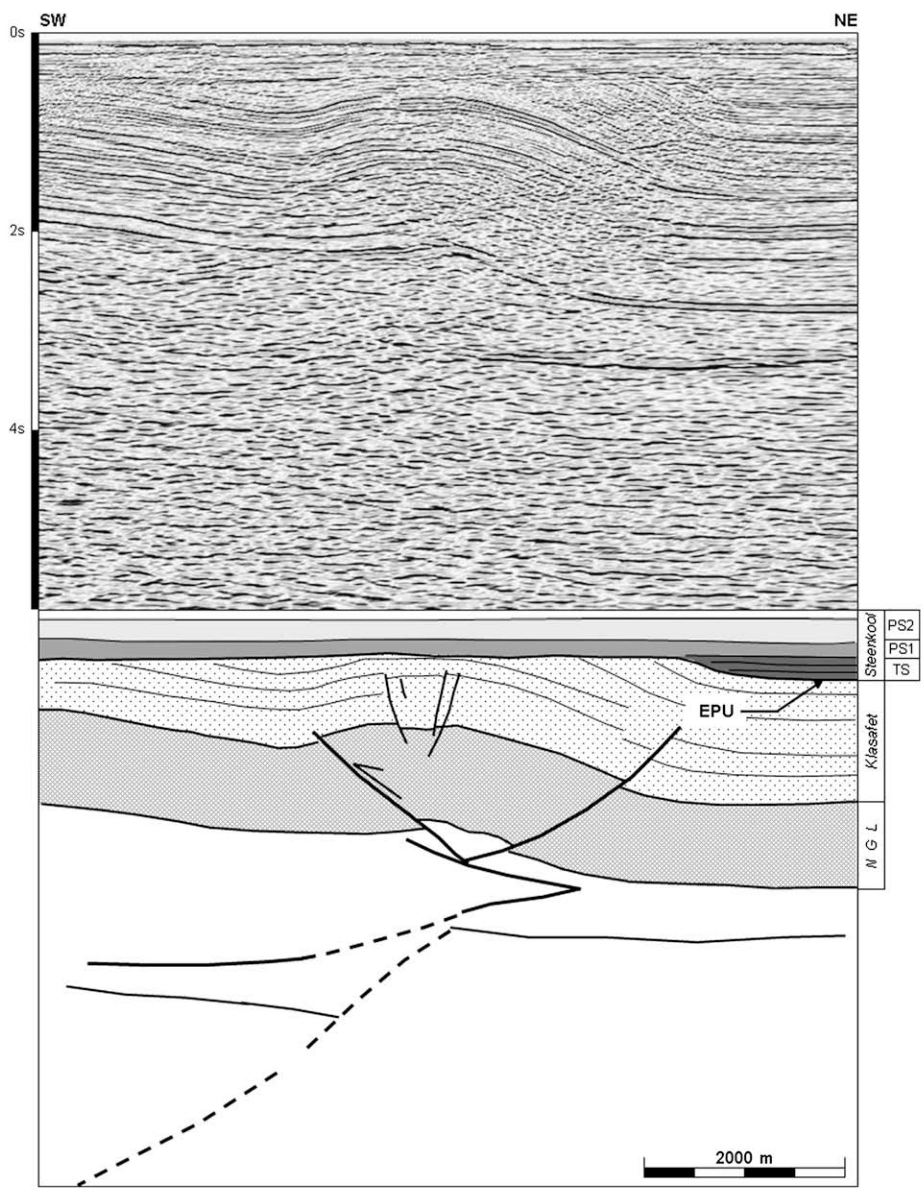

Fig. 6. Pliocene ramp anticline over a normal fault (1:1 scale, location on Figs. 3 and 5). This fold, created during the Late Pliocene, is younger than the most of the folds comprising the MOKR area (Fig. 7) but older than the recent folds in the Seram Trough. It may represent the transition between the early thin-skinned expression of the Seram wedge and the thick-skinned formation of the MOKR during the Late Pliocene.

4.2. Misool Island area: MOKR interaction with the Sorong fault

The Salawati Basin is along the left-lateral Sorong fault, in the northwestern part of the study area (Fig. 4). It is at the geometric position of releasing bend (ENE–WSW to NE–SW) of the Sorong fault. The Sorong fault has been an active left-lateral fault since the Miocene (Dow and Sukamto, 1984). The active motion is demonstrated by focal mechanism (McCaffrey, 1996) and spatial geodesy (Puntodewo et al., 1994; Stevens et al., 2002). The Salawati Basin is a very deep (over 4000 m of sediments) basin where the main normal faults strike NE–SW, and dip to the southeast (Figs. 7 and 8). Some have been inverted or sheared recently (Figs. 7 and 8) resulting in the emergence of the North Misool Archipelago.

Some Pleistocene terraces are seen on the northern coast of the Misool Island. A couple of reasons for their uplift can be considered: one possibility is the existence of a blind thrust beneath this area (Figs. 7 and 9); an alternative explanation is the inversion of the large normal faults of the southeastern margin of the Salawati Basin during the Pleistocene (Figs. 7 and 8).

The southern flank of the MOKR dips gently toward the Seram Trough (~2.5°) and is affected by several short wavelength thrust-folds (Figs. 3 and 7).

Young normal faults affect Pleistocene sediments southwest of Misool Island (Fig. 3). There are two different scales of normal faults (Fig. 7): (1) small gravity driven faults, rooting on the NGL; and (2) large MOKR flexure-associated faults, affecting the whole sedimentary cover (example of the Kumawa Dome, Fig. 3).

4.3. Western "Bird's Head" area: Onin and Kumawa Domes, two topographic anomalies

As in the Misool Island area, a gently dipping southwestern flank is observed in the Onin Dome area. Short wavelength thrust-folds may be seen in the field and on seismic lines (Figs. 3 and 10). In the Kumawa Dome area, the ridge looks more similar to the central area, with a short southwestern flank facing a strong positive Bouguer anomaly located beneath the Seram wedge (Fig. 4).

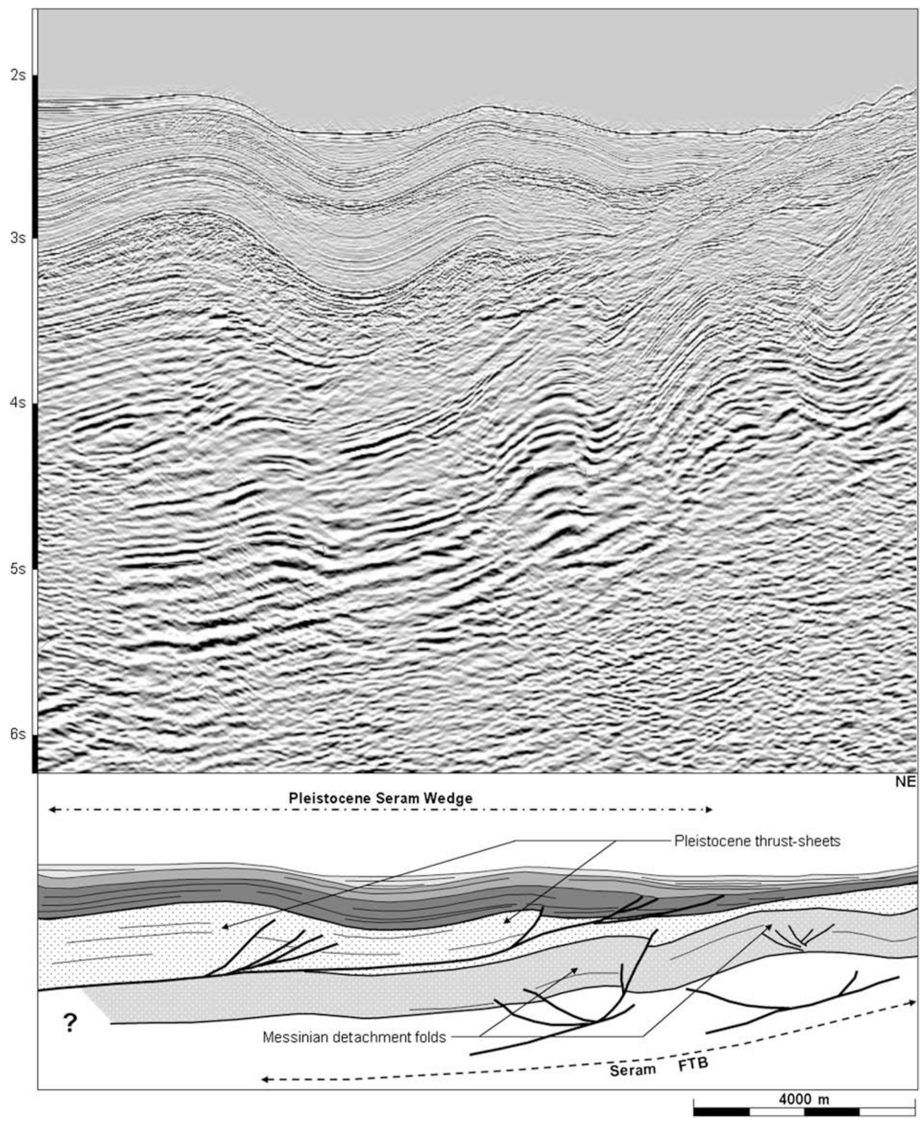

Fig. 7. Early folds and thrusts of the Seram FTB and late Seram wedge thrust-sheets using the southeastern flank of earlier folds as a detachment ramp (bottom: 1:1 scale, location on Fig. 3, colours legend on Fig. 6). The first thin-skinned episode is Late Miocene (~Messinian). These folds and thrusts (Seram FTB) used the Permian–Paleocene Shales unit as detachment layer. The second thin-skinned episode corresponds to the Seram wedge activity resuming by the Pleistocene. Folds from this second thin-skinned episode use the earlier ones as detachment ramps and the clayey Klasafet Fm. as the detachment level.

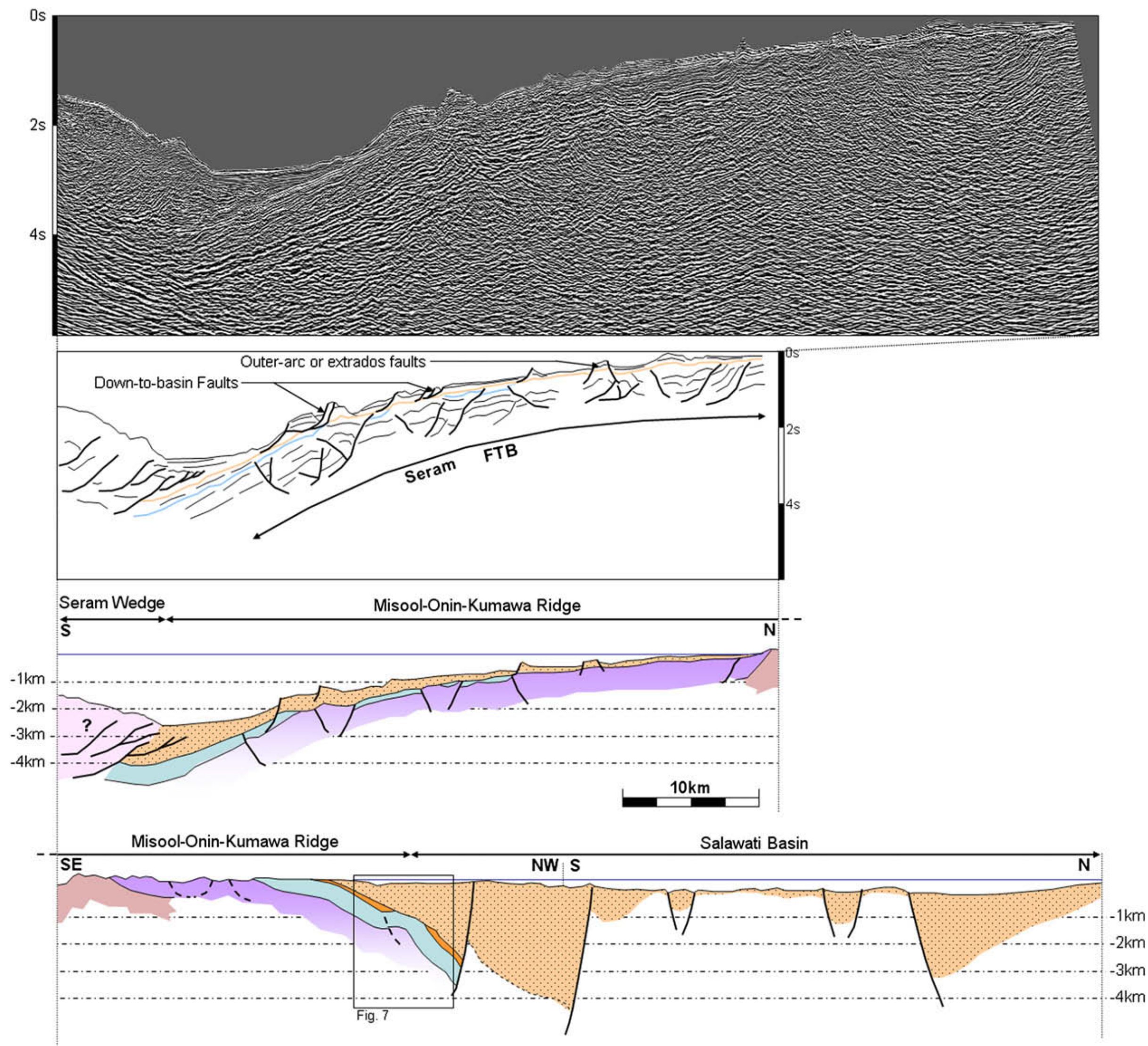

Fig. 8. Cross-section 1: seismic interpretation (top) extended to a MOKR cross-section in the Salawati Basin (bottom, vertical exaggeration x2, location on Fig. 3 and legend on Fig. 5). The Seram wedge appears poorly active in this area. As in the Onin Dome area, the southwestern flank is gently dipping toward the trough. This flank is affected by small normal faults. The emerging Plio–Pleistocene terraces on the northern coast of Misool Island may indicate the presence of a blind thrust (Fig. 10) under the northern coast of Misool Island. In the Salawati Basin, the major normal faults have been recently inverted. The inversion may be the consequence of the Seram wedge reaching the slope of the MOKR.

Along the MOKR from Misool Island to Kumawa Dome the only active structures are: (1) the frontal thrusts of the Seram wedge (Figs. 3 and 11); (2) the normal faults on the southwestern flank of the MOKR (south Misool Island and west Kumawa Dome, Fig. 4); and (3) the Salawati Basin inverted normal faults appear to have been recently active (Fig. 8), although they do not show on the seismicity catalogs (Engdahl et al., 1998).

5. Timing of the deformation

Correlations between seismic reflectors and unconformities identified in wells constrain the timing of tectonic events. The wells used in this study are located on Fig. 3. The age brackets (Fig. 12) are given by well dating (based on palynology and marine fossil biostratigraphy).

5.1. The Seram FTB, Messinian short wavelength folding episode

A thin-skinned and diffuse compressional episode is responsible for the formation of most of the short wavelength folds in the area (Fig. 3). In places where the Klasafet Fm. is entirely preserved (south and southwest Onin Dome), it is possible to evaluate the timing of their formation (Figs. 11 and 13) by using the biostratigraphic ages given by the well South-Onin_1. Folding seems to be contemporaneous with the deposition of the lower part of the Klasafet Fm. (Figs. 11–13). The detachment level is likely to be located within the Permian–Paleocene Shales (Fig. 11). This episode is contemporaneous with the Lengguru FTB formation (Bailly et al., 2008; Fig. 12) at the beginning of the Messinian (seismic correlation with the South-Onin_1 well; Audley-Charles et al., 1979). It represents the first thin-skinned episode (Fig. 12).

This episode is associated with the Seram subduction system because of the structural trend, mainly NW–SE, parallel to the Seram Trough.

5.2. The Sorong fault and the formation of the Salawati Basin

The Salawati Basin is at the geometric position of a releasing bend of the left-lateral Sorong fault (Fig. 3). Seismic correlation

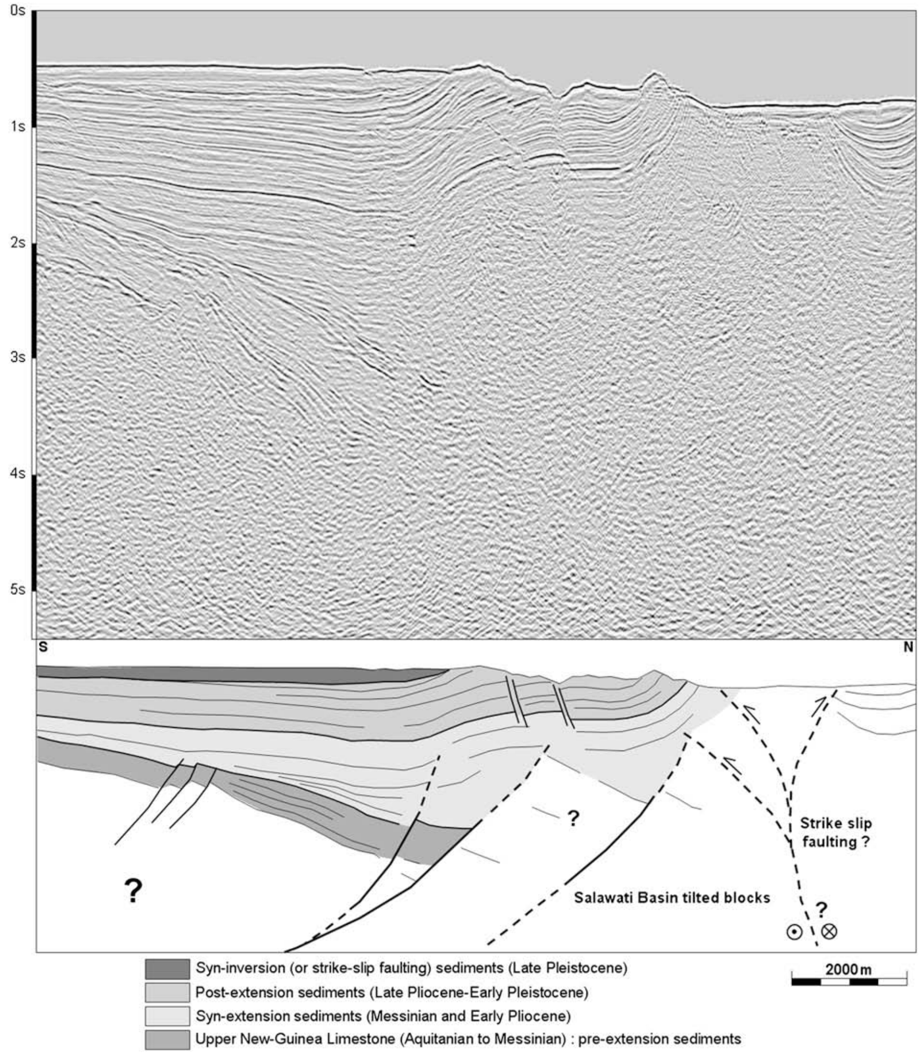

Fig. 9. Detail of the inversion of the Salawati Basin normal faults possible by transpressive tectonics (bottom interpretation at 1:1 scale, location on Fig. 3). Large NE–SW, southeast dipping, normal faults of Late-Miocene–Pliocene times in the Salawati Basin are inverted very recently (Pleistocene).

with the TBFn_1 well show that active extension in the Salawati Basin is inferred to have started in the late Miocene (also directly correlated from Misool Island outcrops), which is consistent with the age given by Pigram and Panggabean (1981).

The sequences and structures observable on seismic lines indicate the Salawati Basin was formed in three stages (Figs. 8 and 14): (1) an extensional regime concentrated on the southeastern basin edge (large normal faults); (2) migration of the deformation toward the centre of the basin with occurrences of minor normal faults; and (3) recent inversion of the main normal faults on southeastern basin edge.

5.3. The formation of the MOKR by the end of the Pliocene

According to Pairault et al. (2003), the Steenkool Fm. may be divided into three sequences of which only the first is folded (Pairault et al., 2003; this study, Fig. 15). This first sequence is approximately the same age as the long wavelength folding and is overlain by Upper Pliocene sedimentary rocks which were later folded during deformation interpreted by Pairault et al. (2003) to be due to development of a forebulge north of Seram Trough. This implies about 2–3 km of uplift (total erosion of NGL Gp.) with an uplift rate between 1 and 2 mm/year and a rapid (less than 2 Myr) construction of topography during the Pliocene.

With the exception of the ramp-fold on the northeastern front of the MOKR (Fig. 6), which is Late Pliocene (Fig. 6), none of the short wavelength folds affect the first sequence (Fig. 15); meaning it postdates the first episode of folding. These different episodes mark two different stages of deformation both in time and wavelength (Fig. 12). The second episode may indicate a thick-skinned deformation (Fig. 12)

As noticed by Pairault et al. (2003), the two distinct folding/bulging episodes appear to control the difference between the core of the MOKR (where the oldest sediments crop out or lay directly beneath the EPU (Fig. 3)) and the overall trace of the MOKR (axis of the Late Pliocene deformation, Fig. 3).

5.4. New thin-skinned deformation; the recent frontal part of the Seram wedge

Large thrust-sheets and folds are observed in the Seram Trough (Fig. 11). They compose the recently active frontal part of the Seram wedge. They have been active since the Pleistocene, the Klasafet and Steenkool Fms. being deformed (Fig. 11) and syn-tectonic layers

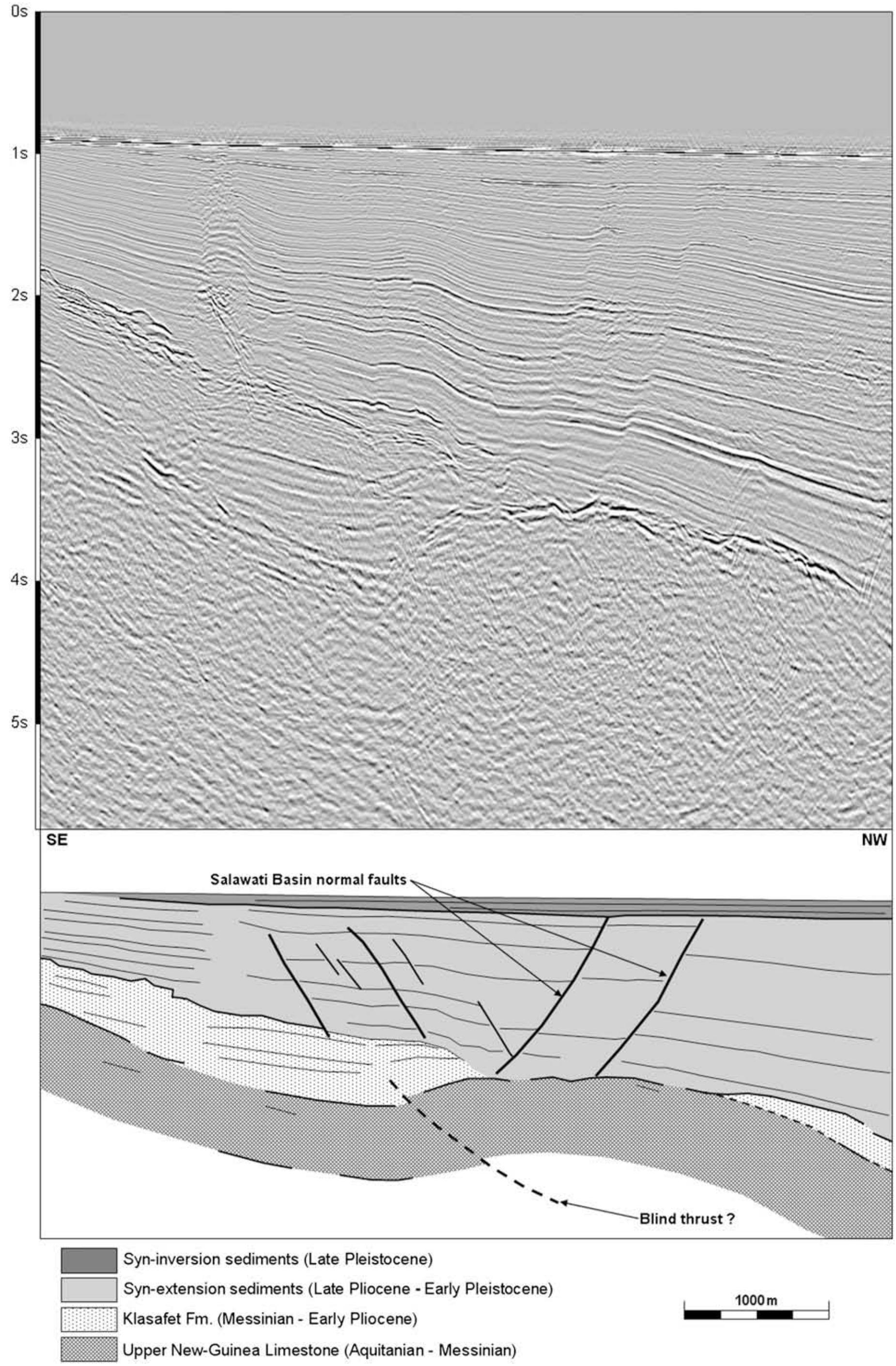

Fig. 10. Detail of the blind thrust under the southeastern margin of the Salawati Basin (bottom interpretation at 1:1 scale, location on Figs. 3 and 7). This blind thrust may have been recently (Pleistocene) reactivated causing the emergence of the Pleistocene terraces along the northern coast of Misool Island.

being located near the top of the Steenkool Fm. (upper part of the second and third sequence of Pairault et al., 2003). Older rocks on Seram Island, including mainly Triassic to Mid-Miocene sediments (Hamilton, 1979), may have served as a backstop for this new wedge which therefore marks a second episode of thin-skinned tectonics (Fig. 12).

The present detachment layer is located at the base of the Klasafet Fm., just above the New Guinea Limestone (Fig. 11). The wedge does not appear to propagate far onto the BH microplate, and was possibly blocked at the location of the former Late Miocene shelf edge. It may therefore post-date the ridge uplift and the formation of the Lengguru FTB (Pairault et al., 2003; Bailly et al., 2008; this study).

The weight of the Seram wedge may have possible effects on the MOKR. We believe that loading is responsible for the normal faults on the southwestern MOKR flank in west Kumawa Dome and southwest Misool Island (Figs. 4, 7 and 13). In addition, the well-marked unconformity resulting from the ridge uplift is buried and dips regularly toward the present Seram wedge (Pairault et al., 2003; this study).

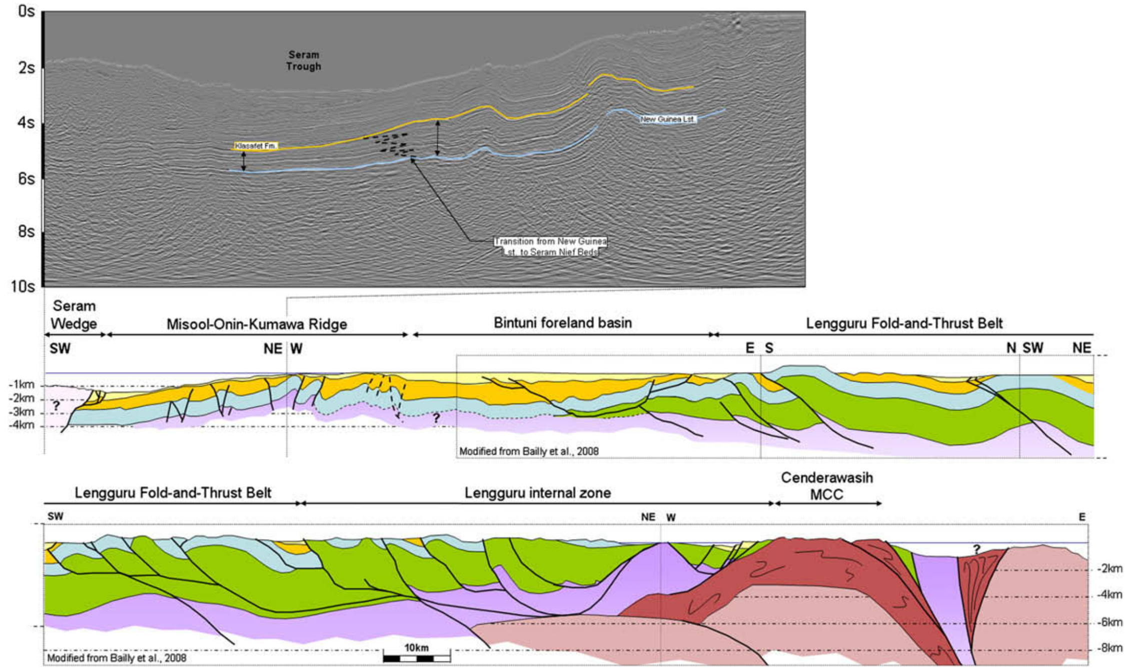

Fig. 11. Cross-section 3: from the Seram wedge to the Lengguru FTB: two facing wedge fronts (vertical exaggeration x2, location on Fig. 3 and legend on Fig. 5). Note the large bulge of the MOKR between the two wedges.

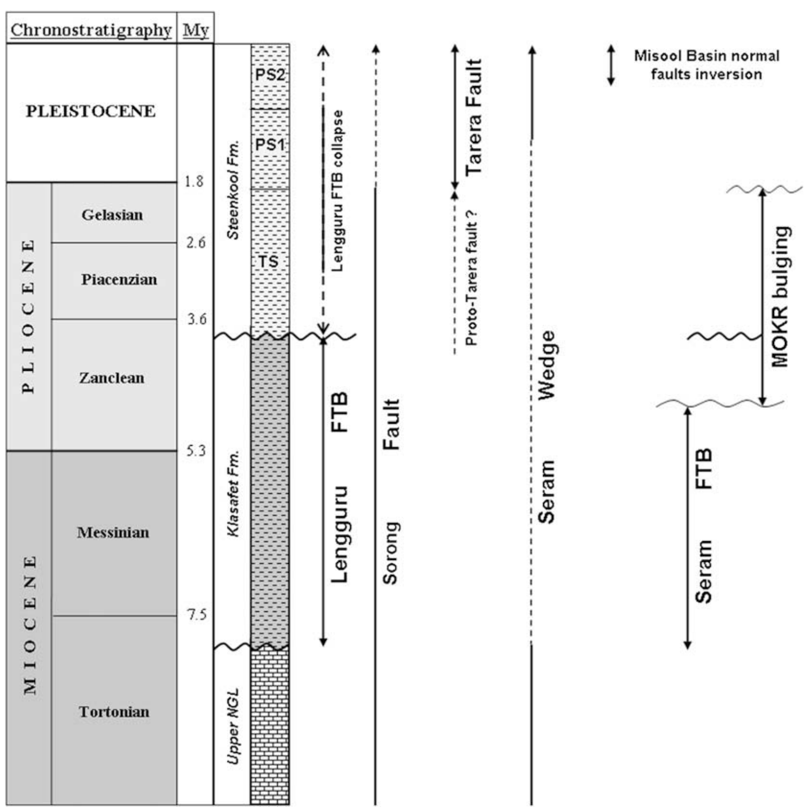

Fig. 12. ''Bird's Head'' microplate tectonic summary. Formation ages are from Pairault et al. (2003) and controlled by wells. The Seram FTB and the MOKR seem to be structures accommodating the convergence between BH microplate and Seram Island instead of the Seram wedge in the latest Miocene and Pliocene. The Seram wedge activity (sensu-stricto) resumed recently and is contemporaneous with the recent Tarera fault (Pubellier and Ego, 2002) which relays the Sorong fault in time (Stevens et al., 2002). Note that these three episodes form a continuum (without cessation of shortening) and represent jumps in the deformation front. The Lengguru FTB, not currently active and collapsing, was fully active by the beginning of the formation of the MOKR.

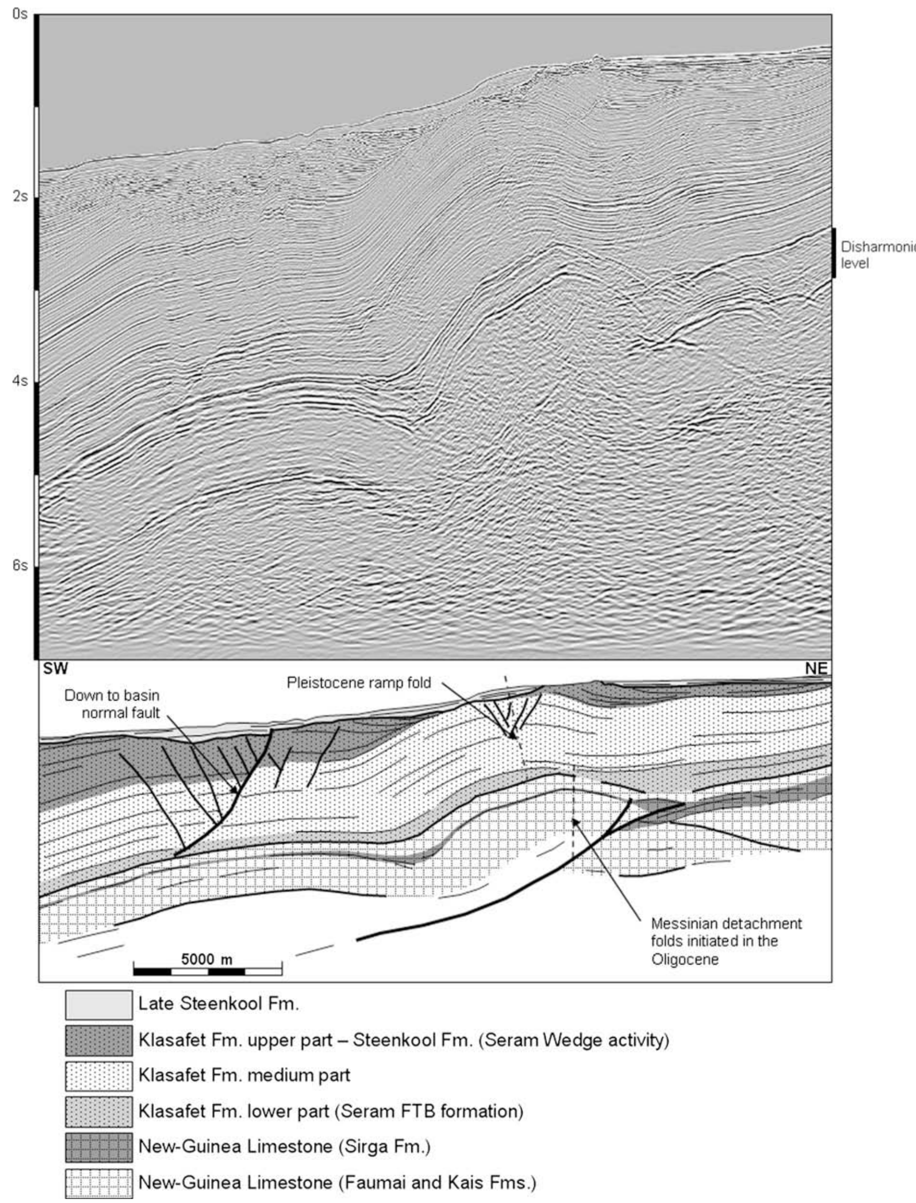

Fig. 13. Early folds and thrusts of the Seram FTB and late down-to-basin normal faults using the southeastern flank of earlier folds as a detachment ramp (bottom: 1:1 scale, location on Fig. 3). This seismic line shows four stages of deformation. The first one, in the Oligocene, affecting the Sirga Fm. is very minor and not usual here. It may be correlated to the Oligocene obduction in New Guinea (Thery et al., 1999). The second and third stages are also illustrated on Fig. 9 and represent the two thin-skinned episodes in the MOKR formation. The latest episode corresponds to down-to-basin normal faulting, perhaps due to the movement of the Seram wedge on the southwestern flank of the MOKR.

6. Discussion: a crustal sheet model

Because the MOKR has a long wavelength and is therefore a crustal or lithospheric scale structure and considering its specific internal structures, the youngest deformation on the MOKR (Pliocene) cannot be interpreted as a result of a simple forebulge effect related to either Seram Trough (as pointed out by Pairault et al., 2003) or the Lengguru FTB. Firstly, its important elevation (above 1000 m in the southern part of the Kumawa anticline) is not in agreement with lithospheric flexure studies (Bieniawski, 1967; Levitt and Sandwell, 1995; Watts, 2001; Turcotte and Schubert, 2002); secondly the variations of the width in the MOKR (50 km in the central area to 150 km in the Misool Island and Onin Dome areas) and the distance of its axis from the Seram Trough (Pairault et al., 2003) suggest a different tectonic process. According to studies of fold wavelength versus decollement depth (Sherwin and Chapple, 1968; Davis et al., 1983; Chester et al., 1991; Fischer and Woodward, 1992; Marshak and Wilkerson, 1992; Doglioni and Prosser, 1997; Teixell and Koyi, 2003), we consider that the latest Late Pliocene deformation on the MOKR is due to a northeastward-moving crustal sheet carried over a detachment located at the top of the ductile crust (about 12 km deep, Fig. 16). Its northeastward vergence is suggested by the continuity of the Late Pliocene northeastern structure (south-westward ramp-fold, Figs. 3 and 6).

In the central part of the MOKR, the short flank is reversed, possibly as a result of the effect of the formation of back-thrusts accommodating more shortening than the main thrust (Fig. 16). In front of the inverted zones, the presence of high Bouguer anomalies beneath the Seram wedge (white arrows, Fig. 4) may provide a possible explanation if interpreted as indicators of crustal asperities of the subducted plate beneath the Seram wedge,

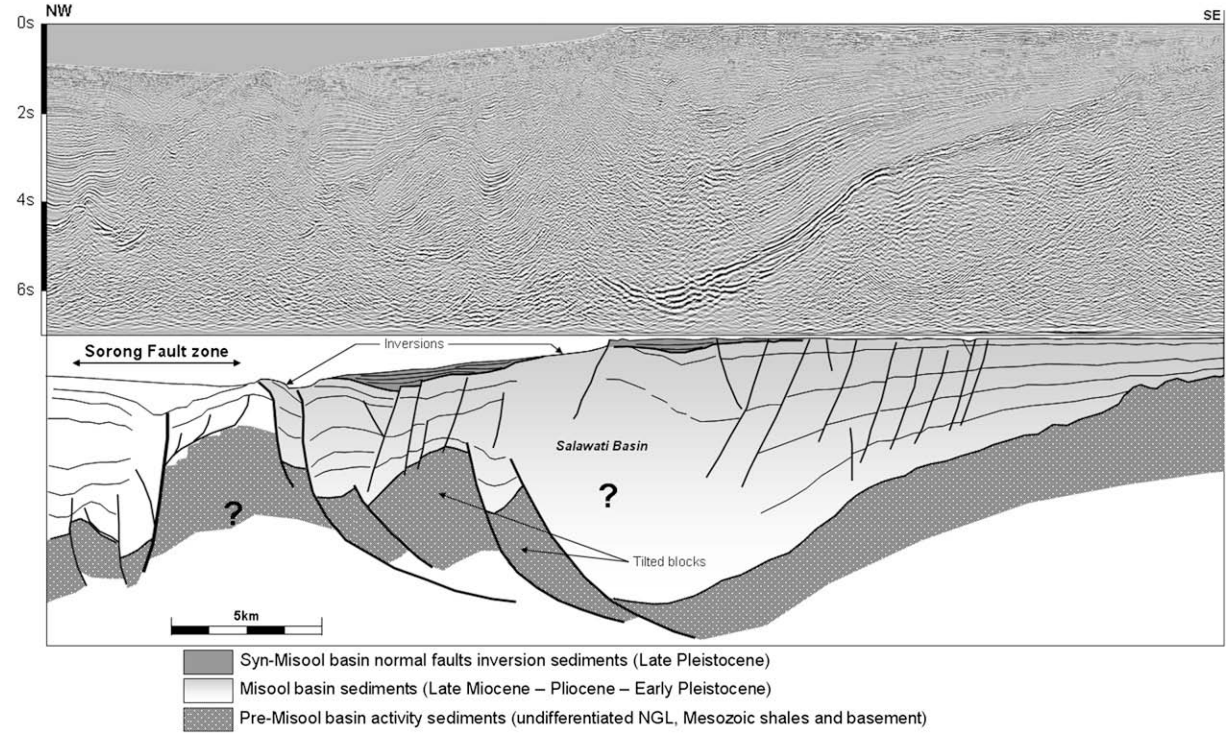

Fig. 14. Salawati Basin cross-section (approx 1.5 vertical exaggeration, location on Fig. 3). The Salawati Basin was an active deep (more than 4 km) basin in geometric relation with the Sorong fault. The Sorong fault is curved in the N of the basin and it is quite probable that the Salawati Basin is a releasing bend of the Sorong fault which has a left-lateral displacement. The basin is composed of two kinds of SW–NE trending faults: (1) large normal southeast dipping listric faults, affecting the basement; and (2) small scale normal northwest dipping faults, only affecting the young sediments.

increasing the rugosity on the subduction/detachment surface in these areas.

The overall geometry represented by thin and thick-skinned thrusts with opposite vergencies is actually that of triangle zone. Examples of such structures are common in front of wedges (e.g. Alberta, Bolivia, etc.). This area would thus represent a triangle zone involving brittle crust and sediments.

To explain the formation of the MOKR in this particular geodynamic setting of double convergence, we favour a model in which continuous convergence on the Seram subduction system is

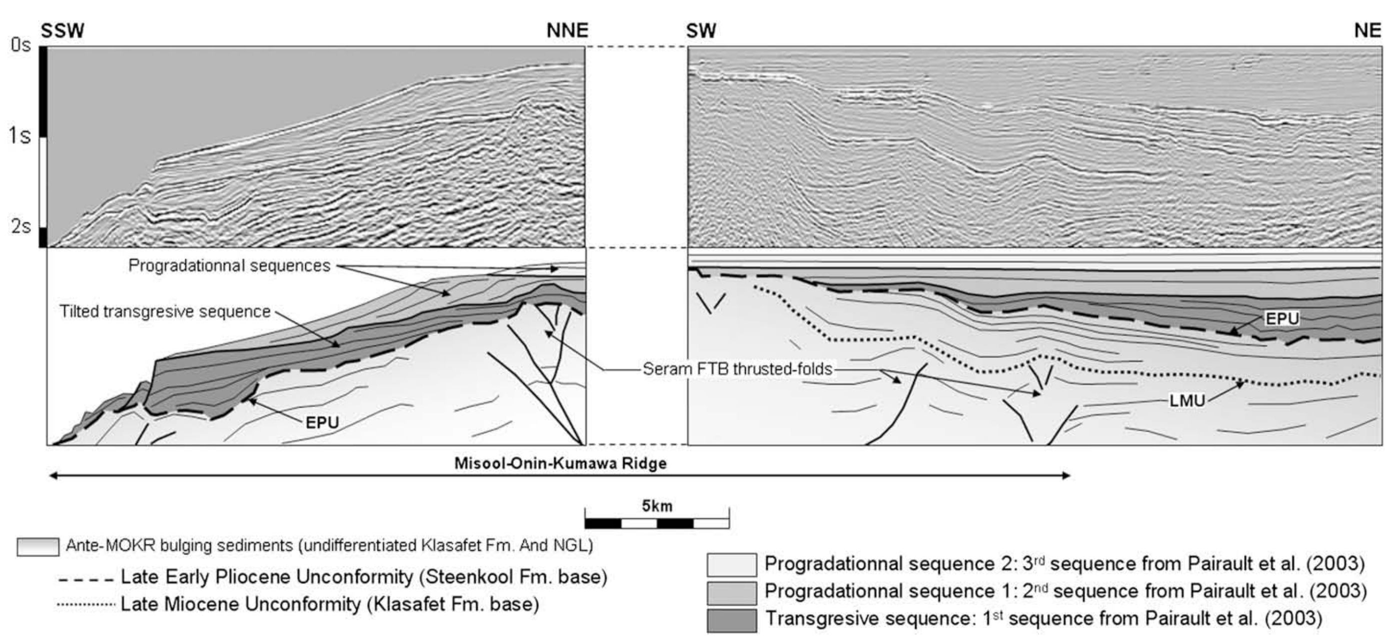

Fig. 15. Overview of the MOKR (approx 1:2 scale, location on Fig. 3). Above the EPU, note the deposition of three sequences, all parts of the Steenkool Fm. (Pairault et al., 2003). The first sequence is slightly deformed by a large scale bulge that folds the unconformity surface that formed the MOKR (Pairault et al., 2003, this study). Dated as Late Pliocene, this sequence is an evidence of the end of the MOKR large scale folding.

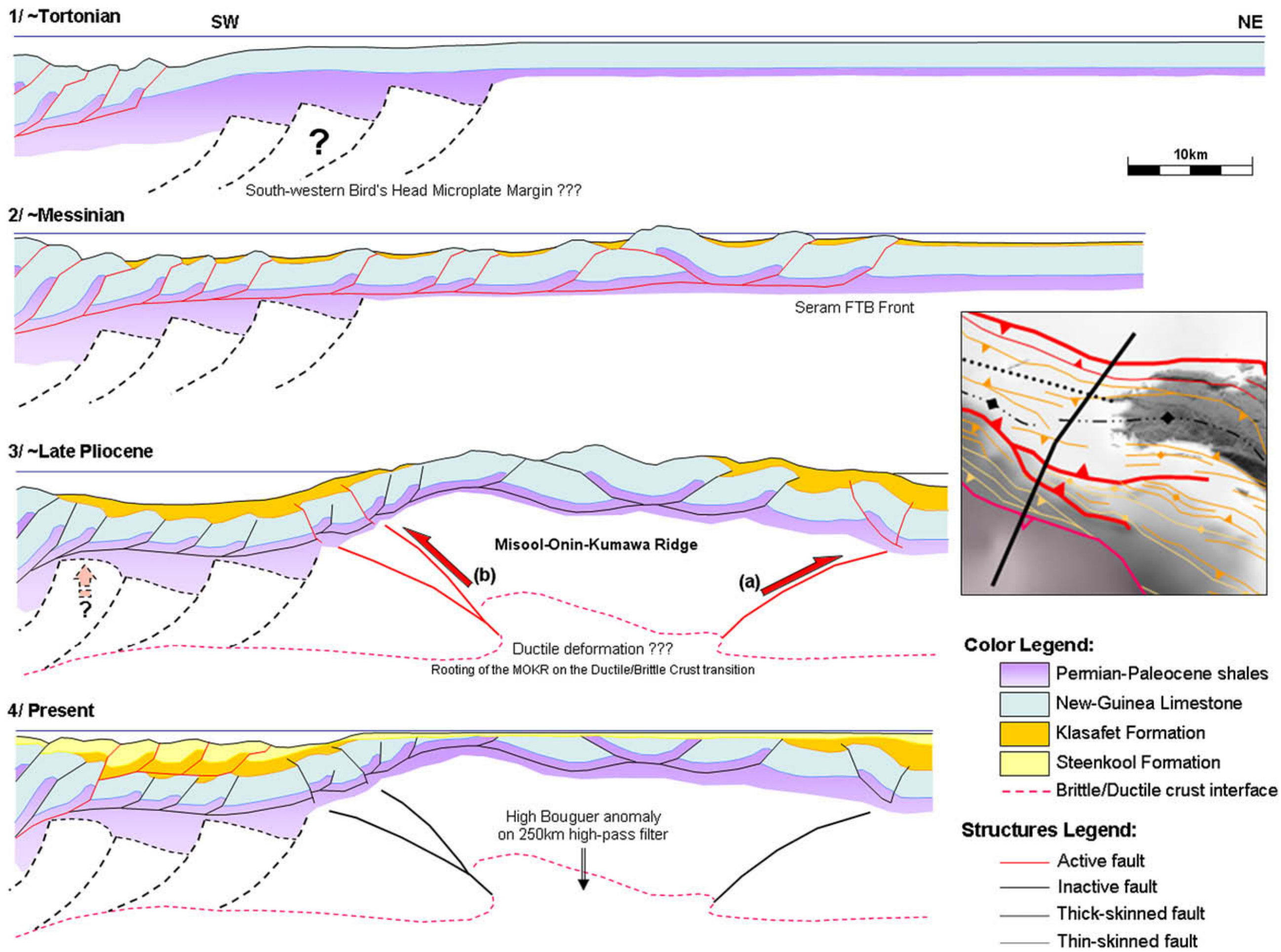

Fig. 16. Geodynamic evolution and formation of the MOKR through the very end of the Tertiary. (1) ~Tortonian: since the Eocene, a large carbonate platform developed. (2) ~Messinian: an early thin-skinned episode resulted in the formation of a FTB far ahead from the Seram wedge. The Permian–Paleocene Shales unit serves as detachment level. (3) ~Late Pliocene: thick-skinned episode leading to the formation of the MOKR likely in response to the subduction of the ancient margin, marking the beginning of continental subduction. A large thrust-sheet (a), northeastward, takes place in front of the Seram wedge, probably flattening at the brittle/ductile crust interface. A thin-skinned effect of this crustal deformation is the formation of a kink fold with an opposite vergence on the northeastern flank of the ridge. In some areas (central MOKR and west Kumawa Dome), we observe the formation of back-thrusts (b) accommodating more deformation, possibly due to the presence of small-scale asperities of the subducting margin (uplifted blocks, horsts?) (4) Recent thin-skinned tectonic resumes only in the Seram Trough. The ancient wedge is used as a "backstop" for the new one.

expressed in different ways through time. We document several local events due to changes of the decoupling levels (Fig. 16):

(1) Until the very end of Miocene, carbonate sedimentation is little disturbed. Tectonic structures are restricted to relics of the Oligocene bulging (Pigram and Symonds, 1991; Thery et al., 1999) such as folds in south Onin (Fig. 13) that developed during the deposition of the Sirga Fm.
(2) During Messinian times (Figs. 11 and 13; coherent with Audley-Charles et al. (1979) Late Miocene deformation dated on Seram Island), the effect of the deformation in the Seram area is expressed far ahead of the deep crustal plate boundary, and shortening constructs the early Seram FTB. The deformation front is located far from the main convergent boundary. This situation is not uncommon and exists in the foreland basin of the Longmen Chan in Sichuan (China). The building of the early FTB may be facilitated by the thickening of the more incompetent Permian–Paleocene Shales.
(3) During Late Pliocene times (Fig. 15), a crustal thrust in front of the Seram Trough began to develop creating topography on the MOKR. This crustal deformation episode is genetically linked to the Seram subduction system because of its shape and vergence. The non-parallelism between the Seram wedge and the MOKR trace (Pairault et al., 2003; Fig. 3) may also be explained by this hypothesis if we consider that the crustal thrusting is facilitated by the presence of crustal discontinuities (such as former sutures or tilted blocks of a former margin). This crustal thrusting episode may be a consequence of the change in the subduction parameters such as subduction of a large scale asperity in the active subduction zone, as the convergence between BH and Seram Island remains very active. Similarly, in central MOKR and Kumawa Dome areas, the ridge is interpreted as a crustal pop-up, due to the subduction of small scale asperities suggested by abnormally high Bouguer anomalies beneath the Seram wedge.
(4) From Pleistocene to present, shortening on Seram wedge has accelerated due to the development of the Tarera fault (Fig. 3, Stevens et al., 2002; Pubellier and Ego, 2002). As a result, the thin-skinned deformation resumed, affecting only young sediments (Klasafet and Steenkool Fms.). The fast convergence may have reactivated some structures of the Seram FTB as ramp faults.

Normal fault inversion in Salawati Basin may result from the beginning of the subduction of the MOKR in the Seram Trough in this area, possibly causing a jump of the deformation front to the northwest of Misool Island.

7. Conclusion

The evolution of the structures around the MOKR took place within a short time span of about 8 Myr, and demonstrates the relative speed at which geodynamic changes occur. The Onin and Kumawa Domes form a major topographic feature in front of the Lengguru FTB, and may represent a local forebulge effect of the Lengguru FTB. The MOKR, in itself, resembles more a thrust of crustal scale. The recent evolution of the MOKR goes through several stages all included in a continuous Neogene convergence between two crustal blocks, the BH and Seram Island (Banda plate). During this evolution, a jump in both the location of the deformation front and the structural level of detachment are observed.

(1) During the Messinian (Audley-Charles et al., 1979; this study): formation of the Seram FTB in front of an active Seram wedge, far ahead of the major crustal convergence boundary. This thin-skinned detachment is responsible for the formation of the majority of short wavelength folds in the studied area.
(2) By the Pliocene: a thick-skinned tectonic style occurs with the formation of a crustal thrust directed northeastward, in response to changes of the Seram subduction parameters (maybe caused by a large scale asperity subduction and/or the evolution of the convergence from an oceanic to a continental subduction, at least from a thinned crust to thicker continental crust). The deformation front does not change from the previous episode but the decollement level cuts deeper, from the Permian–Paleocene Shales (~3 km depth) to the brittle/ductile crust interface (~10–12 km depth).
(3) Recently, thin-skinned deformation resumed principally in the Seram Trough. The decollement level is shallow (Klasafet Fm. base). In places the MOKR begins to subduct, as in south Misool Island area, and the deformation front moves forward into the Salawati Basin.

Such an evolution in three distinct stages even on short time period (~8 My) suggests that the rapid convergence rate (~5 cm/y) between the BH microplate and Seram Island, forced the boundaries to jump both in space and depth. These observations may illustrate the high fragility and instability of deformation zones in fold-and-thrust belts, which is difficult to appreciate in old mountain belts.

In oceanic domains, thick-skinned tectonics are rarely invoked directly and the oceanic crust normally constitutes the backstop of the convergence area or acts as a guide for the thin-skinned deformation. In contrast, in fast geodynamic settings where jumps of deformation take place, the basement is immediately involved. The implication of thick-skinned tectonics in the basement of the subducted plate may be an explanation for the shape and the high topography in the MOKR area. This kind of mechanism has been already invoked by Breton et al. (2004) to explain the large amount of deformation and high pressure–low temperature metamorphism of the Jabal Akhdar and Saih Hatat windows in the Oman Mountains. In that particular case, the high convergence between the Arabian and Eurasian plates in the Oman area led to intra-continental subduction (Breton et al., 2004).

In this paper, we tried to demonstrate the importance of crustal layering in a simplified mechanical model of the MOKR formation. This simplified example is built with three ductile layers (lower crust/Permian–Paleocene Shales unit/Late Miocene–Plio–Pleistocene Clastics unit) and two competent/brittle layers (brittle crust/NGL unit). Our final goal is to integrate this evolution in 2D or 2½D models in order to estimate the variations on subduction parameters such as rugosity or wedge thickness.

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