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Sedimentology of the Melawi and Kentungau Basins, West Kalimantan, Indonesia

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University of Wollongong

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SEDIMENTOLOGY OF THE MELAWI AND KETUNGAU BASINS,
WEST KALIMANTAN, INDONESIA
VOLUME ONE

A thesis submitted in (partial) fulfilment of the requirements for the award of the degree of

DOCTOR OF PHILOSOPHY

from

THE UNIVERSITY OF

WOLLONGONG

by

RACHMAT HERYANTO SUTJIPTO
(Ir ITB Bandung, MSc Wollongong Uni.)

Department of Geology
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The contents of this thesis are the results of original research and the material included has not been submitted for a higher degree to any other university or similar institution.

Rachmat Heryanto Sutjipto
ABSTRACT

The Melawi and Ketungau Basins are located in West Kalimantan, Indonesia. The Ketungau Basin developed between a Tertiary subduction complex (Lubuk Antu Melange) on the Kalimantan-Sarawak border and the Semitau High to the south. The basin is about 50 km wide, 150 km long and it continues eastward into the Mandai Basin. The Semitau High is a linear structural zone comprising submarine slope deposits, a belt of Cretaceous subduction complex (Boyan Melange) and Permian granitoid and metamorphic rocks. The Melawi Basin sequences were deposited between the Semitau High and the continental basement (Schwaner Zone) to the south. The basin is about 75 km wide and 300 km long.

The shallow marine to terrestrial sequences in the Melawi and Ketungau Basins were deposited during the Late Eocene to Oligocene. The Melawi Basin succession comprises four main units separated by periods of uplift and erosion: (1) Ingar Formation - a deep outer shelf marine mudstone; (2) Suwang Group - fluviatile sandstone and lagoonal to marine trough shale; (3) Melawi Group - fluviatile to shallow marine clastic deposits; and (4) Kapuas Group - fluviatile sandstone at the top of the succession. The Ketungau Basin consists of a conformable sequence (the Merakai Group) of shallow marine and floodplain deposits, overlain successively by fluviatile sandstone and floodplain to marginal marine mudstone. The alternation between marine and terrestrial sequences, and the presence of three unconformities in the Melawi Basin,
indicates tectonic instability during the depositional histories of the basins.

On the basis of sandstone petrology, diagenesis and depositional facies the Melawi Group and Alat Sandstone in the Melawi Basin can be correlated, respectively, with the Kantu Formation and Tutoop Sandstone in the Ketungau Basin. Both palaeocurrent and provenance studies indicate derivation of the Melawi and Ketungau Basin sequences from the north, mainly from uplifted recycled orogenic material in the Boyan and Lubok Antu Melanges. A few units in the Melawi Basin contain magmatic arc detritus derived from the Schwaner Mountains to the south.

Although both basins contain coal seams, the best quality coal is at the top of the Melawi Basin sequence. Organic maturation and vitrinite abundance indicate that both basins have potential for the generation and entrapment of petroleum.

Late Cretaceous subduction in northwestern Kalimantan deformed the Late Cretaceous marine sequence producing the Boyan Melange which incorporated Permian granitic microcontinental fragments. Uplift of the Simitau High (Boyan Melange) along backthrusts during the Paleocene and Early Eocene produced an accretionary prism flanked to the south by the forearc Melawi Basin. Periodic backthrusting resulted in folding, uplift and unconformities in the northern Melawi Basin. Northward migration of the Benioff Zone in the Late Eocene created the forearc Ketungau Basin between the old and new (Lubok Antu Melange) outer arc ridges.
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CHAPTER 1
INTRODUCTION

1.1 BACKGROUND OF STUDY

The Melawi and Ketungau Basins of West Kalimantan contain thick sedimentary sequences, deposited predominantly in shallow marine and continental environments. The Indonesia-Australia Geological Mapping Project (IAGMP) has carried out systematic 1:250,000 scale geological mapping of a significant part of the Melawi and Ketungau Basins as a part of the production of geological and geophysical maps of both West and East Kalimantan between 1983 and 1988. The author was a member of this project and was partly responsible for the geological map of the Sintang Quadrangle.

The IAGMP operates under an agreement between the two countries. The Indonesian government is represented by the Geological Research and Development Centre and the Australian government is represented by the Bureau of Mineral Resources. As part of the training aspect of this project, Indonesian geologists are given an opportunity to study at a university in Australia. The author is one of the Indonesian geologists who has been chosen by the project to take part in the overseas study program.

1.2 LOCATION

The Melawi and Ketungau Basins are located in West
Kalimantan, Indonesia (Fig. 1.1). The studied area lies in the central part of the administrative province of West Kalimantan, between longitudes 110°30'E and 112°30'E, and between 1°0'N and 0°45'S latitudes. It lies about 400 km east of Pontianak, the capital city of the province, and encompasses the middle reaches of the Kapuas River drainage system. Sintang is the largest town in the studied area and lies at the confluence of the Kapuas and Melawi Rivers (Fig. 1.2).

1.3 PHYSIOGRAPHY

The most extensive regions in the studied area consist of the low-lying Kapuas and Melawi River valleys. The former valley is bounded to the southeast by rugged country, strongly dissected by large rivers rising in the Madi Highland. Low-lying swampy ground of the upper Kapuas Lake District lies north of the river, whereas low rolling country extends to the western part of the area. The northern margin of the studied area includes a range of mountains up to 1220 m high which marks the international border with the Malaysian state of Sarawak. The Melawi Valley is bounded to the northeast by the Madi Highland and to the south by the rugged Schwaner Mountains.

The studied area can be divided into seven topographic zones (Fig. 1.2) which can be related to the underlying geology (Fig. 1.3).
1.3.1 Lake District Depression

The northeastern part of the studied area is covered by large lakes which are generally shallow and variable in extent from year to year. The lakes are inter-connected by perennial waterways. They have formed in a recent depression and are isolated from the main river system by broad low ridges composed of either older alluvial deposits or recent levee deposits. In the Lake District two areas of resistant, granitic rock are present forming substantial hills.

1.3.2 Upper Kapuas Meander Belt

The meander belt extends south from the low ridges of older alluvium forming the southern margin of the Lake District Depression to the foothills of the Madi Highlands. The region is flat and swampy, and becomes partially inundated during wet season flooding. Many oxbow lakes, cut-off river sections and mud-filled oxbows are present. The meander belt continues west to the edge of the Lake District Depression close to the Tawang River junction.

1.3.3 Ketungau Hills

A range of hills rising to 1220 m occurs along the northern margin of the studied area. Two types of topography occur in this region. The first consists of northwest trending ridges composed dominantly of resistant quartz sandstone (Tutoop Sandstone), with dip slopes facing southwest and steep back-slope scarps, e.g. Mt Tutoop. The second topographic type is a group of inselbergs composed of
basaltic to granodioritic high-level intrusive rocks.

1.3.4 Central Topographic High

A line of low hills of altered mafic volcanic rocks occurs sporadically in the Boyan Melange and continues west to a group of hills composed of Permian granitoid and metamorphic rocks. The hills extend from Mt Condong in the center of the studied area to Mt Bengkawan and Mt Besi in the west. The surrounding countryside consists of a low-lying but dissected landscape with abundant linear features on aerial photographs which represents faults and joints.

1.3.5 Melawi and Ketungau Peneplains

Most of the studied area is covered by low rolling hill country with a characteristic dendritic drainage pattern. It is underlain by sedimentary sequences of the Melawi and Ketungau Basins. Around and east of Sintang, inselbergs caused by resistant igneous intrusions, are common on the peneplain (e.g. Mt Kelam).

1.3.6 Madi Highland

This constitutes an upland region composed of many different landforms related to geological features. Long linear ridges in the Silat River area and in the far southeast are caused by steeply dipping sandstone ridges which control the drainage direction. Large, gently sloping areas of low relative relief are underlain by gently dipping quartz sandstone. These areas occasionally form plateaux
(e.g. Mt Alat and Mt Merdja).

1.3.7 **Schwaner Mountains**

The southern part the studied area contains a group of mountains and hills composed of tonalite, granodiorite and granite batholiths with minor mafic rocks and low grade regional metamorphic rocks.

1.4 **CLIMATE, VEGETATION AND POPULATION**

The climate of the studied area is equatorial, and the monsoonal influence is only minor. The average annual rainfall at Sintang is 3528 mm. There is no great rainfall variation across the area, and no clearly defined wet and dry season, although a lower rainfall and more dry days are usually recorded between June and September.

Large parts of the studied area are covered by secondary rain-forest, but areas in the southeast and north support primary rain-forest which has become the target for intense logging activity. Logs are transported in vast rafts along the Kapuas and Melawi Rivers to saw-mills.

Population in the studied area is widespread with small villages and longhouses usually located adjacent to rivers navigable by outboard-powered canoes. Several large towns are located at major river junctions. The town of Sintang is the largest with a population of about 20,000 people. Most of the population is engaged in traditional agriculture, with growing of hill-rice and corn occupying the majority of the agricultural effort. Cash crops including rubber, tapioca
(cassava), citronella and pepper are grown, the latter two only in a small way. A small number of people are engaged in mining alluvial gold, commonly in conjunction with subsistence farming.

1.5 PREVIOUS GEOLOGICAL WORK

Prior to 1939 all the major geological work in West Kalimantan had been carried out by the Dutch East-Indies mining office "Dienst van het Mijnwezen in Nederlandsch Oost-Indie". The earliest expedition to Kalimantan was led by Schwaner (1844-1847), and it was followed with expeditions by Everwijn (1853-1857) and Van Schelle (1879-1886). These expeditions provided the first descriptions of the geology and geography of the region, and in particular provided the first collection of fossil material for European workers. An expedition in the upper Kapuas region from 1893 to 1895 by Molengraaf (1900) provided invaluable geological data. The results of these investigations were reported in the yearbooks (Jaarboek van het Mijnwezen in Nederlandsch-Indie) from 1872 onwards. The first geological synthesis was written by Easton (1904), whereas a geological map (1:100,000 scale) of West Kalimantan was prepared by Van Es (1918). Rutten (1927) prepared a lecture series on the geology of the East Indies.

Van Emmichoven supervised field work in West Kalimantan in the late 1920's and early 1930's, and the results were reported in the yearbooks and were presented as a 1:250,000 scale geological map of part of West Kalimantan. He wrote
several important reports on the geology up till the late 1930's.

Van Bemmelen (1949) used all of these data and summarized them into a general geological model. This geological model was expanded on by Haile (1969, 1974). Hamilton (1979) provided a geological synthesis of West Kalimantan as part of a geological and structural synthesis of Indonesia.

Geological surveys have also been carried out in adjacent Sarawak, including reports by Tan (1979, 1982 & 1986), Sengor (1984, 1986) and Wood (1985).

The Indonesia-Australia Geological Mapping Project carried out systematic 1:250,000 scale geological and geophysical mapping of the whole area of West and East Kalimantan between 1983 and 1988.

Several oil companies also carried out preliminary exploration programs, including geological, seismic and gravity surveys, during recent times. Unfortunately the results of these works are confidential.

1.6 GEOLOGICAL BACKGROUND

1.6.1 Regional Geology

Kalimantan is the largest island in Southeast Asia and occupies a central position in this area. Consequently the geology of Kalimantan provides information that can be used towards understanding the geology and tectonic development of the Southeast Asian region.

Van Bemmelen (1949) discussed the geology of Kalimantan
in terms of the geosynclinal theory, with a geosyncline extending as an arcuate belt across Kalimantan. Haile (1969, 1974) discussed the tectonic history in terms of both geosynclinal and plate tectonic theories, and he maintained the concept of continuous arcuate structural belts across Kalimantan.

A subduction model for the geology of Sarawak was proposed by Tan (1979, 1982), and he introduced the concept of a regionally extensive melange belt in the southern part of Sarawak. The complex geology in the region north of the melange belt was interpreted by Hamilton (1979) as a huge accretionary wedge. He also suggested that an earlier Jurassic subduction event was recorded in Kalimantan. Sengor (1984, 1986) suggested that the ophiolite in West Kalimantan occurred on a possible pre-Late Jurassic thrust, while Wood (1985) considered the belt of pre-Tertiary deformed rocks in West Sarawak and West Kalimantan to be an uplifted pre-Cretaceous tectonic unit (structural high) on the trans-Kalimantan shear zone. Williams et al. (1986) considered that the Boyan Melange in West Kalimantan is of Late Cretaceous age and is not related to any Jurassic event. Williams and Heryanto (1986) argued that Cretaceous mass flow deposits and melange are part of a Late Cretaceous accretionary wedge and not remnants of a coherent Palaeozoic craton as previously proposed by Van Emmichoven (1939).

Williams et al. (1988) identified three main geological domains in West Kalimantan which are overlain or onlapped by extensive sedimentary basins. The stratigraphic correlation
of each domain is shown in Figure 1.4. They interpreted the Late Cretaceous to Eocene tectonic history of West Kalimantan as an accretionary event, involving construction and deformation of a sedimentary accretionary wedge which now lies on continental basement. They also argued that no continuous arcuate structural zones extend right across northwest Kalimantan.

1.6.2 Local Geology

The Melawi and Ketungau Basins are separated from each other by a structural "high" (Semitau High) formed of older rocks comprising submarine slope deposits and a belt of Cretaceous subduction complex (Boyan Melange; Fig. 1.3).

The Melawi sequences were deposited within a basin developed in a linear zone between the Semitau High to the north and the continental basement (Schwaner Zone) to the south. The Tertiary Melawi sequences are composed of open marine mudstone at the base, overlain by fluviatile sandstone and lacustrine shale in the middle, and shallow marine clastic and fluviatile sandstone at the top of the succession. This basin contains over 7500 m of clastic sediments. It is an elongated basin, about 75 km wide, which extends eastward across West Kalimantan for over 300 km from basement near the west coast to a possible junction with the Kutai and Barito Basins in the central part of Kalimantan.

The Ketungau succession was deposited in a basin developed between the Semitau High and a Tertiary subduction complex to the north. The Late Tertiary Ketungau succession comprises
shallow marine clastic sequences, overlain by fluviatile sandstone which is covered by shallow marine to lacustrine mudstone. The elongate Ketungau Basin is about 50 km wide and 150 km long and trends eastward into the Mandai Basin.

Both Tertiary sequences were intruded by basaltic to rhyolitic intrusive rocks, commencing in the Late Oligocene in the south and progressively extending northward during the Early Miocene.

Finally most of the rocks in the studied area are covered by Quaternary deposits.

1.7 AIM OF THE STUDY

The various sandstone formations within the Melawi and Ketungau Basins were defined during the IAGMP, but these formations look physically similar to each other. The basins have been surveyed by several authors since the 1800's, but each author provided a different opinion about the formations, geology and tectonic settings.

The aim of the present study is to define significant similarities and differences between the various sandstone formations within both the Melawi and Ketungau Basins and to determine the depositional environments, petrography, provenance, diagenesis and resource potential of each stratigraphic unit. Additional aims are to use these data to correlate the two basin sequences and to determine the tectonic implications of the basin development in relation to the geological history of northwest Kalimantan.
CHAPTER 2
GEOLOGICAL SETTING

2.1 REGIONAL GEOLOGY

A geological map of the studied area is shown in Figure 2.1, and representative stratigraphic columns are presented in Figure 2.2. The oldest rocks in the studied area are Permian granitoid and metamorphic rocks which crop out in the northwestern part of the area. Blocks of these granitoid and metamorphic rocks are also caught up in the Boyan Melange which is of Cretaceous age. K-Ar dating on samples of both granitoid and metamorphic rocks collected by the IAGMP yielded minimum ages of 320 Ma. The granitic rock contains a strong foliation and is associated with schist, phyllite and quartzite which show a garnet grade greenschist facies assemblage.

The granitoid and metamorphic unit is overlain by a Late Triassic continental to shallow marine unit. The equivalent rocks in Sarawak contain detritus from the older granitic suite (Tan, 1986), implying an unconformable relationship. The Triassic unit contains Monotis and Halobia (Williams et al., 1988). Early Jurassic ammonites and bivalves were identified by Easton (1904) in shallow marine shale calcareous and nodular siltstone, which is intercalated with feldspathic conglomerate, and biohermal, oolitic and intraclastic
limestone. Williams et al. (1988) suggested that the Early Jurassic beds appear to be conformable with the Late Triassic strata.

Late Jurassic clastic near-shore detritus and shallow marine limestone forms a marginal facies to a north-trending trough which is dominantly filled with Cretaceous sandy turbidite beds and calcareous mudstone (Williams et al., 1988). Tan (1986) noted that an hiatus occurred between the Late Triassic strata and the Late Jurassic-Cretaceous rocks in Sarawak as indicated by a structural divergence between these units.

The southern part of the studied area is occupied by the Schwaner Batholith which consists of tonalite, granodiorite and granite with minor amounts of low grade metamorphic rocks. Radiometric age determinations on this batholith yield dates ranging from Jurassic (157 Ma) to Late Cretaceous (77 Ma; Haile et al., 1977). Recent K-Ar age determinations carried out by IAGMP yielded only Cretaceous ages (129 Ma to 87 Ma).

The Semitau High between the Melawi and Ketungau Basins comprises the Selangkai Formation and the Boyan Melange. The Selangkai Formation is composed dominantly of calcareous mudstone with intercalations of boulder and pebbly mudstone, graded sandstone and rare limestone and conglomerate. This formation is lithologically variable and some stratigraphic levels are dominated by turbidite sandstone or other mass flow deposits. The thickness of this formation is difficult to estimate because of the presence of tight folding and the lack
of a clear stratigraphic succession. Van Emmichoven (1939) suggested a total thickness of more than 3250 m, and Williams and Heryanto (1986) agreed that a total thickness in excess of 3000 m may be realistic.

Thirty samples from the sandstone and mudstone within the Selangkai Formation have been analyzed for microfossils by Dr A.R. Lloyd; 3 of them are barren and some of them contain very rare foraminifers. Eleven samples yielded Foraminifera indicating a Turonian age, based on the presence of Lenticulina spp., Nodosaria spp., Eponides diversus, Heterohelix globulosa, Heterohelix striata, Globigerinelloides aspera, Globotruncanina linneiana, Rotalipora sp. cf. R. greenhornensis, Saracenaria sp. From 9 samples of limestone blocks and lenses, 7 of them yielded Orbitolina scutum indicating a Cenomanian age and 2 samples yielded radiolarians indicating a Cretaceous age (Williams and Heryanto, 1986). Previous determinations of fossils from this formation (listed by Van Emmichoven, 1939) include ammonites, bivalves and foraminifers which yielded an Early Cretaceous (Valanginian) to Late Cretaceous (Turonian) age.

The sedimentary structures found in this formation are sole marks, parallel lamination and common ripple marks. Locally mass flow deposits, olistostromes, open framework immature conglomerate and turbidite sandstone also occur in the sequence. The appearance and variety of rock types and sedimentary structures suggest that the rocks were deposited by submarine gravity flows in a basin accumulating pelagic and hemi-pelagic calcareous mud. The facies associations ranging
from boulder beds to thin laminated mudstone suggest that a submarine fan is a suitable depositional environment for this formation.

The Selangkai Formation is strongly deformed. Bedding is, in places, overturned and two fold phases have been identified locally. The main fold axis is horizontal, trending to 280°, but folding is asymmetrical with the southerly dip predominant and steeper than northerly dips. Generally the stratigraphic sequences are younger to the north, which indicates a large-scale overturned or imbricated structure in the unit. South-dipping thrust faults were recorded by Williams and Heryanto (1985).

Two zones of melange have been defined in the area (Fig. 2.1) by Williams and Heryanto (1985). The southern zone is termed the Boyan Melange while the northern zone in Sarawak was termed the Lubok Antu Melange by Tan (1979, 1982).

The Boyan Melange is a multiply deformed polymict tectonic breccia composed of fragments and blocks of a wide variety of sedimentary, metamorphic and igneous rocks in a highly cleaved and pervasively sheared pelitic matrix. The fragments and blocks are mostly angular, although some are subangular or irregular, and they range in size from a few centimetres to many kilometres. Fragments comprise sandstone, mudstone, shale, limestone, conglomerate, chert, schist, quartzite, granodiorite, granite, dacite, diorite, dolerite, gabbro, basalt and serpentinite. The largest blocks, up to 6 km across and 40 km long, are composed of mafic and intermediate rocks and their metamorphosed equivalents. The matrix is dark
grey or dark brown-grey and weathers to a mottled light grey and light yellow. It is pervasively sheared with a lustrous surface giving it a scaly appearance. It is non-calcareous, practically barren of fossils and is cleaved and fractured.

Fossils have been identified in 7 limestone samples from the melange. They include *Orbitolina scutum*, molluscs, echinoids, foraminifers, corals, radiolarians and indeterminate fossil fragments. The presence of *Orbitolina scutum* suggests a Cenomanian age for the clasts and, therefore, the Boyan Melange is probably younger than Cenomanian. To date no Tertiary fossils have been found in the melange.

The Lubok Antu Melange in Sarawak was described by Tan (1982) as a unit of chaotically distributed blocks of chert, limestone and igneous rocks in a pervasively sheared matrix. It ranges in age from Late Cretaceous to Eocene. Poor outcrops of this unit have been found in the studied area.

The Melawi and Ketungau Basin sequences consist of fluviatile, lagoonal and marginal marine sediments, with the Ketungau Basin sequence probably being slightly younger than the Melawi Basin sequence (Fig. 2.3). The basins are separated from each other by the Semitau High consisting of the Selangkai Formation and Boyan Melange.

Intrusions of the Sintang basaltic to rhyolitic high-level hypabyssal porphyritic rocks have yielded ages ranging from 23.0 Ma to 30.4 Ma for sample from the southern part of the Sintang area, and 16.4 Ma to 17.9 Ma for samples from the northern part (Williams and Heryanto, 1986). These data
indicate that magmatic activity commenced in the southern part of the study area in Late Oligocene and continued to Early Miocene time, extending progressively northwards in the Early Miocene.

Most of the rocks which occur in the study area are partly covered by Quaternary deposits.

2.2 MELAWI BASIN

The Melawi Basin is a structural depression containing up to 7500m of fluviatile, lagoonal and marginal marine sedimentary rocks. These sequences are strongly deformed in the middle part of the basin and gently deformed in the remainder. They crop out between the Schwaner Mountains in the south and a strongly deformed sequence of submarine fan deposits to the north (Fig. 2.1). The Melawi Basin is 75 km wide and extends eastward for over 300 km from basement near the west coast to a possible junction with the Kutai or Barito Basins in central Kalimantan (Fig. 2.5).

A modified schematic cross-section based on seismic interpretation of line WM 108 (Total Oil Co., 1986) in the Kedukul area, western Melawi Basin, can be seen in Figure 2.6. Sketch geological cross-sections for the eastern Melawi Basin modified from seismic interpretations from Elf Aquitaine Indonesie (B. Porthault, pers. comm., 1987) are shown in Figs. 2.7 and 2.8.

Based on its structural appearance, the Melawi Basin sequence can be divided into one formation and three groups (Fig. 2.3) as follows: Ingar Formation, and Suwang, Melawi and
Kapuas Groups. A composite stratigraphic column for the Melawi Basin is shown in Figure 2.9.

2.2.1 **Ingar Formation**

The term *Ingar Facies* was introduced by Van Emmichoven (1939) as a possible lateral facies equivalent of the Lebang Mudstone in the Tertiary Melawi Beds. His argument for this correlation was firstly that the structural style of the Ingar Facies is similar to that of the Melawi Beds, and secondly that the Ingar Facies changed gradually into Lebang Mudstone which is, in turn, conformably overlain by plateau-forming sandstone. Van Bemmelen (1949) subsequently included this formation in his Melawi Beds or Melawi Facies. Marks (1957) described the Ingar Facies as a lateral equivalent of the Lebang Claystone Member at the top of the Melawi Beds.

Some oil companies involved in hydrocarbon exploration in West Kalimantan have suggested that the following three Eocene formations underlie the Silat Fold Belt: the Mangan Formation overlain by the Mentomoi Shale and the Ingar Formation (B. Porthault, pers. comm., 1987; Table 2.1).

Generally the rocks of the Ingar Formation are gently folded. Local zones with intense shearing sub-parallel to bedding are associated with tight recumbent to inclined folds. The eastern and southern margins of the the Ingar Formation are unconformably overlain by the Suwang and Melawi Groups respectively. The northern boundary with the Selangkai Formation is probably a fault.

The Ingar Formation occurs in the central part of the
studied area cropping out along the Ingar River about 50 km east of Sintang. It extends west to the Sanggau area and south to the Nangapinoh area. The formation forms a low undulating topography and supports a dendritic drainage pattern. Outcrops are sparse and in a few places the formation is covered by swamp.

The Ingar Formation consists of grey calcareous mudstone interbedded with siltstone and fine-grained sandstone. The mudstone locally contains sandstone lenses and calcareous concretions ranging from 5 cm up to 1 m in diameter (Fig. 2.10). The interbedded sandstone is grey, very fine- to fine-grained and slightly to strongly calcareous. Framework grains consist of quartz and feldspar, some rock fragments and mica.

The thickness of the Ingar Formation is not known since both the top and bottom of the formation have not been seen in one section. A thickness of at least 2000 m is estimated from the outcrop width and regional dip at the Ingar River section.

The base of the Ingar Formation has not been observed. Aerial photograph interpretation and ground observation indicate that the relationship between the Ingar Formation and the Selangkai Formation to the north is a fault. Contacts with the overlying Dangkan Sandstone to the east and Payak Formation to the south are unconformable.

2.2.2 Suwang Group

Williams and Heryanto (1985) introduced the term Suwang Group for the Silat Fold Belt sequence which includes the
Dangkan Sandstone and Silat Shale.

The Suwang Group occurs as a narrow (15 km wide) belt which extends eastwards for over 150 km from the Kapuas River to the Madi Highland in the central part of the study area. It unconformably overlies the older submarine fan deposits to the north. The Silat Fold Belt was folded before the deposition of the Melawi Group and is unconformably overlain by the Melawi and Kapuas Groups.

2.2.2a DANGKAN SANDSTONE

The name Dangkan Sandstone was introduced by Williams and Heryanto (1985) for the tightly folded sandstone forming a narrow ridge in the Silat River area, Sintang Quadrangle. The name was derived from the administrative name for the middle reaches of Silat River.

The unit was first described by Easton (1904) as a steeply dipping complex of rather thinly interbedded hard and soft sandstone with intercalated mudstone lying on the Cretaceous sequences. Loth (1918) described the steeply dipping sandstone at the Dangkan River, a tributary of the Silat River, and Van Es (1918) called this unit the Plate Sandstone which he considered to be older than the Plateau Sandstone. Molengraaf (1900), Van Emmichoven (1939), Van Bemmelen (1949) and Marks (1957), however, included this sandstone within the Plateau Sandstone.

The Dangkan Sandstone forms a prominent continuous double ridge which outlines the Silat Syncline. The northern limb extends from the Kapuas River eastwards for over 100 km to the
Madi Highland while the southern limb changes direction from eastward to southward around Mt Bang. The Dangkan Sandstone consists of a polymict open framework basal conglomeratic coarse-grained sandstone overlain by lithic arenite.

The basal conglomeratic coarse sandstone is white to brown and contains abundant sub-rounded to rounded quartzite pebbles in a ferruginous quartz-rich matrix. The sandstone is well compacted and is massive to thickly bedded (3-5 m).

Lithic arenite is the dominant rock type in the formation. It is white to brown in colour and ranges from medium-grained to pebbly sandstone. Framework grains consist of quartz, quartzite and chert, with minor schist and mafic minerals. The average bed thickness is 30 - 50 cm (Fig. 2.11) and ranges up to 2 m.

The Dangkan Sandstone unconformably overlies the Selangkai and Ingar Formations, whereas the contact with the overlying Silat Shale is conformable. The thickness of the Dangkan Sandstone is estimated to be 600 m, while the largest measured section at Nanga Ngeri village, in the middle reach of the Silat River, was 530 m (Appendix 2).

2.2.2b SILAT SHALE

The Silat Shale was first described by Easton (1904) as a thinly foliated, crumbling shale in the mid-course region of the Silat River. Icke and Martin (1905) noted that the Silat Shale is older than Eocene but younger than Cenomanian, and they considered it to be probably of Late Cretaceous age, which harmonized very well with its tectonic relations. Van
Es (1918) ascribed a Late Cretaceous age to the shale, and he considered it to be the youngest division of Cretaceous rocks in the region. Van Emmichoven (1939) described these dark shales as the Silat Group which he considered to overlie the "Plateau Sandstone" with a transitional boundary. However, Van Bemmelen (1949) described the Silat Facies in the Semitau Zone as a large lens of claystone in the Plateau Sandstone, formed by deposition in a long lagoon. Marks (1957) classified the Silat Shale as a member of the Silat Group and considered the age to be Paleogene.

The Silat Shale occupies the core of the Silat Syncline and crops out along the Silat River, extending eastwards for over 100 km from the Kapuas River to the Madi Highland. The Silat Shale consists of black, strongly-sheared and slickensided (Fig. 2.12), carbonaceous shale with some turbidite sandstone beds. It contains a rare molluscan fauna. Lenses of very fine-grained micrite and thin interbeds of siltstone are commonly found within the slaty shale. Bed thickness averages 5 to 10 cm, but ranges up to 30 cm. The siltstone is grey to black and contains some thin coal lenses, plant remains, fragmented carbonaceous material and concretions.

The thickness of the Silat Shale is not known because the upper boundary has not been seen in the field. The thickness is estimated to be 2000 m from the dip and outcrop pattern in the Silat Syncline (near the head of the Silat River). The largest measured section of this formation at Suwangkelik Creek, a tributary of the Suwang River, was 1825 m thick (Appendix 3).
The relationship between the Silat Shale and the underlying Dangkan Sandstone is conformable with the contact, as observed in the field, being gradational. The contact with the overlying Melawi Group is an angular unconformity.

2.2.3 Melawi Group

The Melawi Group was redefined by Williams and Heryanto (1985) to include lagoonal and marginal marine sediments in the southern part of the Melawi Basin, i.e. the Payak and Tebidah Formations. In the southern part of the basin the Melawi Group shows marked lateral facies changes. The western part of the group consists predominantly of sandstone, which passes eastward into interbedded mudstone, siltstone and sandstone with some coal seams. This predominantly sandstone unit is now recognized as the Sepauk Sandstone. The Melawi Group unconformably overlies the Ingar Formation and Suwang Group in the north and onlaps granitic and metamorphic basement (Schwaner Batholith) to the south. It is overlain by the Kapuas Group, and forms an asymmetrical sub-basin, with maximum sediment accumulation and steeper limb dips near the northern margin. The rocks are folded into a gentle asymmetrical syncline with maximum limb dips of 30°.

2.2.3a. SEPAUK SANDSTONE

A thick medium- to coarse-grained sandstone unit forming the lower part of the Melawi Group in the south-central and southwestern part of the basin is now recognized as a separate formation, the Sepauk Sandstone. The unit was originally
described as a facies of the Tebidah Formation but was separated as a distinct formation by petroleum exploration companies (Table 2.1).

The formation starts with a thin conglomerate or breccia lying directly on granitic basement. The overlying sequence is medium- to coarse-grained with a few pebbly sandstone lenses. Medium scale (17-23 cm thick) planar tabular and trough cross-beds are common in the sandstone beds. Scour surfaces are common in between the sandstone beds. The lower part of the sandstone occurs as poorly defined fining-upwards sequences and is interpreted to represent fluvial channel deposits, whereas the upper part may represent nearshore marine sands.

The Sepauk Sandstone unconformably overlies the basement rocks of the Schwaner Zone along the southern margin of the basin. The nominated type section is along the Sepauk River (Fig. 3.48) where the Sepauk Sandstone is about 1650 m thick and is conformably overlain in the eastern and central parts of the basin by a shale unit of the Tebidah Formation. In the west it is disconformably overlain by the Sekayam Sandstone. The Sepauk Sandstone is well developed in the western part of the basin but becomes thinner and finer grained eastwards to the Pinoh River, and then rapidly decreases in thickness towards the Serawai River as a result of interdigation with the shales of the Tebidah Formation. To the north the Sepauk Sandstone thins out and probably interfingers with the Payak Formation.
2.2.3b PAYAK FORMATION

The term Payak Formation was introduced by Williams and Heryanto (1985) for the alternating tuffaceous quartzwacke, mudstone and siltstone which crop out in the southern part of the Sintang Quadrangle. Van Es (1918) included these rocks in the Melawi Group; Van Emmichoven (1939) placed the formation in the lower division of the Melawi Unit; while Van Bemmelen (1949) included them in the Melawi Facies. Marks (1957) described rocks belonging this formation as the lower member of the Melawi Group and called them the Melawi Formation.

The Payak Formation trends eastward and southeastward across the central part of the study area as a series of parallel strike ridges. This formation consists predominantly of tuffaceous quartzwacke (Fig. 2.13), alternating with grey mudstone (Fig. 2.14) and siltstone which are richly fossiliferous in places. The quartzwacke is dirty white to light grey, very fine- to fine-grained, rarely medium- to very coarse-grained. It consists dominantly of quartz and feldspar with some mica flakes and carbonaceous material. The unit is well bedded with bed thickness usually from 10 cm to 30 cm, ranging up to 60 cm. The total thickness of the Payak Formation is approximately 1750 m (Appendix 4) as determined from the measured section in the upstream part of the Payak River.

The Payak Formation unconformably overlies the Suwang Group and Ingar Formation. The contact with the overlying Tebidah Formation is conformable, probably with a gradational boundary. The Payak Formation in the northern part of the
basin interfingers with the lower part of the Tebidah Formation farther south, so that the Payak Formation is not exposed in the southern part of the basin. To the southwest the Payak Formation probably interfingers with the Sepauk Sandstone.

2.2.3c TEBIDAH FORMATION

The Tebidah Formation was named by Williams and Heryanto (1985) to include the uppermost unit mapped in the Melawi Group, comprising lithic arenite interbedded with mudstone. The type locality of this formation is in the Tebidah River, a tributary of the Kayan River in the southeast corner of the study area. Van Es (1918) included these rocks with his Melawi Group. Van Emmichoven (1939) restricted this formation to the lower division of the Melawi Unit. Van Bemmelen (1949) included it with his Melawi Layer or Facies while Marks (1957) described this formation as a lower member of the Melawi Group (Table 2.1).

The Tebidah Formation covers most of the southern part of the study area and extends east to the Melawi River. The formation produces low-lying undulating country, in some places covered by swamps. Outcrop is sparse.

The Tebidah Formation consists of lithic arenite (Fig. 2.15) interbedded with green and red mudstone in the upper part (Fig. 2.16) and grey mudstone in the lower part (Fig. 2.17). It contains thin seams of coal. In the southwest, the lower part of the Tebidah Formation is dominated by sandstone.
with thin conglomerate and breccia beds lying directly on the granitic basement.

The Tebidah Formation is conformable on the underlying Payak Formation, probably with a gradational boundary. To the west the lower part of this formation interfingers with the Sepauk Sandstone. It is overlain by the Sekayam Sandstone with slight angular unconformity in the western part, and it is also unconformably overlain by the Alat Sandstone in the eastern part of the basin. The nature of this contact is visible on aerial photographs and radar imagery, particularly when enhanced by image ratio analysis of the four Landsat bands. The total thickness of the Tebidah Formation is approximately 3000 m. The measured section in Mentatal River was 2925 m thick (Appendix 5).

2.2.4 **Kapuas Group**

Kapuas Group is a new term for the sandstone units which unconformably overlie the Melawi Group. The Kapuas Group consists of the Sekayam and Alat Sandstones, which are laterally correlated units. The Sekayam Sandstone crops out in the western part of the studied area, whereas the Alat Sandstone crops out in the eastern part.

2.2.4a **SEKAYAM SANDSTONE**

The Sekayam Sandstone was defined by Williams and Heryanto (1985) as a greenish grey lithic arenite, interbedded with green, reddish and grey mudstone, which is exposed along the Sekayam River, a tributary of the Kapuas River, in the
Sanggau Quadrangle (western part of the study area). Van Emmichoven (1939) and Van Bemmelen (1949) included this unit in the upper part of the Melawi Group.

The Sekayam Sandstone consists of lithic arenite interbedded with mudstone (Fig. 2.18). The sandstone is greenish grey and generally medium- to coarse-grained; in places it ranges up to very coarse-grained or pebbly. Grains are dominantly quartz and rock fragments (basalt and metaquartzite) with minor feldspar, muscovite and mafic minerals. Grains are generally rounded to subrounded and rarely subangular. Sorting is moderate. In places the sandstone has a calcite cement. The sandstone is well bedded with bed thickness averaging 50 cm to 150 cm, ranging up to 250 cm in places. The total thickness of the Sekayam Sandstone is approximately 150 m.

The relationship between the Sekayam Sandstone and the underlying Payak and Tebidah Formations is unconformable. The Sekayam Sandstone is partly covered by Quaternary deposits. Based on its stratigraphic position above the Tebidah Formation, the Sekayam Sandstone is correlated with the Alat Sandstone.

2.2.4b ALAT SANDSTONE

The Alat Sandstone was named by Williams and Heryanto (1985) to include the gently dipping quartz sandstone which forms the extensive highland plateaux of Mt Alat in the Nangapinoh Quadrangle (southeast corner of the study area) and in the eastern part of the Sintang Quadrangle (northeastern
part of the study area). Molengraaf (1900) was the first to describe this quartz sandstone which he placed in the Late Cretaceous and Neogene Plateau Sandstone. Van Es (1918) defined this sandstone as the Neogene Plateau Sandstone whereas Van Emmichoven (1939), Van Bemmelen (1949) and Marks (1957) included several different quartz sandstone units in the Plateau Sandstone.

The Alat Sandstone consists of massive to thickly bedded chert breccia (Fig. 2.19), conglomerate and quartz-lithic arenite (Fig. 2.20) with thin interbeds of mudstone in places. The light grey breccia is only present in the lower part of the unit. The sandstone is well bedded with an average bed thickness of 30 to 50 cm, ranging up to 150 cm.

The total thickness of the Alat Sandstone is approximately 250 m, whereas the measured section at Mt Bunyau (southeast corner of the study area) was 225 m thick (Appendix 6). The sandstone lies unconformably on the Boyan Melange, Selangkai Formation, Dangkan Sandstone and Tebidah Formation.

2.3 KETUNGAU BASIN

The Ketungau Basin contains shallow marine clastic sequences overlain by fluviatile sandstone, which was later covered by shallow marine to lacustrine mudstone. The lowest formation in the Ketungau Basin is similar to the Melawi Group which prompted Van Emmichoven (1939) to correlate the two units. The thick fluviatile sandstone in the middle of the Ketungau Basin sequence is similar to the fluviatile sandstone in the upper part of the Melawi Basin sequence which
Molengraaf (1900), Van Es (1918), Van Emmichoven (1939), Van Bemmelen (1949), Marks (1957) and Tan (1979) called the Plateau Sandstone. Fossils from the lowest exposed rocks in the Ketungau Basin in Sarawak yielded a Late Eocene age (Tan, 1979), similar to the microfloral age for the Payak Formation in the Melawi Basin. Consequently the Ketungau Basin sequence was deposited after deformation of the Silat Fold Belt in the Melawi Basin.

2.3.1 Merakai Group

Williams and Heryanto (1985) introduced the term Merakai Group for the Ketungau Basin sequence and they divided it into the basal Kantu Formation overlain successively by the Tutoop Sandstone and Ketungau Formation. A composite stratigraphic column for the Ketungau Basin is shown in Figure 2.21.

2.3.1a KANTU FORMATION

The Kantu Formation was originally described by Van Emmichoven and Ter Bruggen (1935) as the Kantu Beds, with the type locality in the Kantu River, a tributary of the Empanang River in the northern part of the study area. They correlated the unit with the Melawi Subgroup on the basis of lithology. The continuation of this formation in Sarawak is called the Silantek Formation (Leichti et al., 1960).

The Kantu Formation is poorly exposed in a belt trending eastward along the southern margin of the Ketungau Basin. It is much better exposed along the Kantu and Empanang Rivers (northern margin of the Ketungau Basin) where it forms a belt
of steep hills which widens towards the northwest as it passes into Sarawak. Large lenses of sandstone in the formation form strike ridges ranging from low swells to high and steep hills.

The Kantu Formation consists of interbedded quartzose, feldspatic and micaceous sandstone, carbonaceous siltstone and mudstone (Fig 2.22). A few carbonaceous laminae (1-5 mm thick) are present (Fig. 2.23). Bedding thickness averages 15 cm to 50 cm and ranges up to 150 cm.

The total thickness of the Kantu Formation in the Kantu River area is estimated to be 4000 m from dip measurements, whereas the total thickness from the measured sections at several places in the Kantu area was 3898 m (Appendix 7). The formation could be thinner on the southern margin of the basin. The Kantu Formation unconformably overlies the Boyan Melange in the south and is faulted against the Lubok Antu Melange to the north. The Kantu Formation appears to be conformably overlain by the Tutoop Sandstone.

The intercalation of grey and reddish mudstone, containing carbonaceous laminae or coal seams, within Kantu Formation is similar to the Tebidah Formation in the Melawi Basin. The age of the Kantu and Silantek Formations is Late Eocene to Oligocene (Tan, 1979), which is similar to the age of the Tebidah Formation and consequently they can be correlated.

2.3.1b TUTOOP SANDSTONE

The Tutoop Sandstone (Williams and Heryanto, 1985) comprises quartz sandstone and forms ridges around Mt Tutoop in northern part of the study area. Molengraaf (1900), Van Es...
(1918), Van Emmichoven (1939), Van Bemmelen (1949) and Marks (1957) included this sandstone in the Neogene Plateau Sandstone.

The Tutoop Sandstone crops out around the Ketungau Basin extending northwestward to form the boundary between Sarawak and Indonesia. The Tutoop Sandstone ridges are separated by V-shaped valleys. The ridges are rounded and of high relief including Mt Tutoop at 1220 m.

The Tutoop Sandstone is a massive to thickly bedded quartz-lithic arenite (Fig. 2.24), with some interbedded conglomerate (Fig. 2.25) and mudstone in places. The sandstone is yellowish white to brownish white, medium- to coarse-grained and pebbly in places. Grains are dominantly quartz and lithic fragments (quartzite and chert) with a little feldspar and mica. They are subrounded to rounded with moderate sorting. The cement is silica. Average bed thickness is 1 m to 3 m and ranges up to 5 m. Some conglomerate beds are present and they commonly rest on scoured mudstone surfaces. The fine-grained tuffaceous sandstone contains very angular recrystallized glass shards.

The thickness of the Tutoop Sandstone in the Segubung River is about 1500 m (Appendix 8). It is conformable on the Late Eocene Kantu Formation and is overlain conformably by the Ketungau Formation. The Tutoop Sandstone is lithologically similar to the Alat Sandstone in the upper part of the Melawi Basin sequence.
2.3.1c KETUNGAU FORMATION

The Ketungau Formation was originally named by Van Emmichoven (1939) as the Ketungau Beds, which he correlated with the Silat Shale. He sub-divided this unit into three stages, with poorly defined boundaries. This usage was followed by Van Bemmelen (1949).

The Ketungau Formation is exposed in the northwestern part of the study area along the Ketungau River. It occupies the center of the large synclinal structure in the Ketungau Basin and forms undulating country with low relief and a dendritic drainage pattern.

The Ketungau Formation consists of shallow marine to non-marine sandstone (Fig. 2.26), siltstone and mudstone with thin coal seams in the upper part. The mudstone is grey to black and well bedded with laminae of carbonaceous material. The beds are generally 10 cm to 30 cm thick and range up to 50 cm. The sandstone is light purplish or yellowish grey, carbonaceous and largely fine-grained, but in places it is medium-grained. Grains consist of subangular to subrounded, moderately to well sorted quartz and feldspar. The siltstone is light grey to yellowish grey, well bedded and contains carbonaceous laminae and concretions of iron oxide. In places it is fossiliferous. Coal seams appear as thin (5-15 cm) intercalations, especially in the upper part of the formation.

The thickness of this formation is estimated to be 1500 m, whereas a measured sections along the Sekapat River was 1480 m (Appendix 9). It conformably overlies the Tutoop Sandstone and it is overlain, in part, by Quaternary deposits.
2.4 PALAEONTOLOGICAL AGE CONTROL

2.4.1 Melawi Basin

Tertiary microfaunas are rare in the Melawi and Ketungau Basin sequences. Equivalent units in Sarawak yielded only four fossiliferous samples from the Silantek Formation, none of which contain age diagnostic assemblages (Tan, 1979).

In the study area deposition in the Melawi Basin began with the Ingar Formation. Fossils are rare in the northern part of the Ingar Formation with only two samples yielding *Rotalipora* sp. and radiolarians that indicate a Cretaceous age (A.R. Lloyd, pers. comm., 1984). Six samples from the southern part of the formation yielded abundant small foraminifers, including *Lenticulina* sp., *Dorothia* sp., *Gaudryina* sp., *Heterohelix papula*, *Hedbergella trocoidea*, *Hedbergella deltoiensis*, *Globotruncan* *a marginata*, *Gyroidina* sp. and radiolarians (Appendix 10). These fossils indicate a Late Cretaceous (Turonian) age (A.R. Lloyd, pers. comm., 1984). Based on these fossils Williams and Heryanto (1985, 1986) concluded that the Ingar Formation is Late Cretaceous.

Two samples from the Ingar Formation were examined for microfaunas by Elf Aquitaine Indonesie (B. Porthault, pers. comm., 1987). They contain *Globotruncan* *a* sp., *Heterohelix globulosa* and *Gavelinella* sp. These fossils indicate a Turonian to Santonian age. Thirty seven field samples and two core samples from the Melawi Basin were also studied for calcareous nanoplankton by this foreign oil company (B. Porthault, pers. comm., 1987). Only 8 field samples and 1 core sample were found to contain nanoplankton, which are
scarce and poorly preserved. They contained Watznaueria barnesae, Eiffellithus eximius, Eiffellithus turriseiffeli, Cretarhabdus crenulatus, Zygodiscus diplogrammus, Eprolithus floralis, Prediscosphaera cretacea, Lucianorhabdus cayeuxii, Cribrosphaerella ehrebergi and Nanoconus sp. (Appendix 10) which indicate a Turonian to Santonian age. On the other hand abundant very diversified palynological assemblages were found in the Ingar Formation (from 41 examined samples 33 were fossiliferous) during recent hydrocarbon exploration by the same oil company (B. Porthault, pers. comm., 1987). Many species provide precise stratigraphic assignment and floral assemblages range from very poor to very rich. They include Ephedripites type D (sensu Muller), Psilatricolpites acuticostatus, Triorites festatus, Ephedripites ovalis, Triorites tenuieximus, Psilatricolpites kayanensis, Retitricolpites vulgaris, Pterocarya sp., Alnipollenites sp., Lanagianopollis microrugulatus, Araucariacites australis, Alnipollenites verus, Quercoidites sp., Discolpopollis elegans, Gothanipollis bassensis, Iguanura sp., Classopollis torosus, Victoria sp., Marginopollis concinnus, Striatocolpites conspicuus, Browlonia sp., Caryapollenites sp., Gemmatricolpites pergemmatus, Ilexpollenites sp., Zonocostites ramonae, Florschuetzia trilobata, Sapotaceae, Proxapertites operculatus, Margocolporites vanwijhei, Rhizophora sp., Echiperiporites estelae, Ctenolophonidites costatus, Myrtaceidites sp., Ulmipollenites sp., Proxapertites cursus, Ctenolophonidites lisamae, Praedapollis flexibilis, Polypodiaceoisporites speciosus, Sphagnumsporites sp.
Acrostichum sp., Distaverrusporites margaritatus, Verrucatosporites usmensis, Verrucingulatisporites sp., Distaverrusporites simplex, Cicatricosisporites dorogensis and Magnastriatites howardi (Appendix 10).

Many of the species found within the palynological assemblages provide a precise Eocene age for the Ingar Formation. Reworked older fossils are also present in these assemblages. Thus the microfaunal assemblages found in Ingar Formation must represent reworked Cretaceous material.

The open marine shelf or bathyal depositional environment inferred by Williams and Heryanto (1985) on the basis of Cretaceous microfossils is, therefore, invalid. However, the presence of Dinophyceae may indicate a marine environment (B. Porthault, pers. comm., 1987).

The Eocene Ingar Formation is unconformably overlain by the unfossiliferous Dangkan Sandstone. Deposition of the Dangkan Sandstone was followed by the Silat Shale. Nine samples from the Silat Shale were analyzed for microfossils by IAGMP, but eight of them were barren. One sample contained rare radiolarians. A molluscan fauna of gastropods and bivalves was found within a black shale bed. Icke and Martin (1905) provided a description of the fauna, which they considered to be Late Cretaceous in age. However based on the stratigraphic position of the Dangkan Sandstone and Silat Shale, which unconformably underlie the Late Eocene Payak Formation and overlie the Eocene Ingar Formation, an Eocene age is indicated for both these units.
The Payak Formation unconformably overlies the Silat Shale. Abundant small bivalves have been found in some parts of the Payak Formation, but they have not so far been identified. Five samples from this formation were analysed for microfossils but two of them were barren. Microfossils identified from the Payak Formation include *Rotalia cushmani*, *Globotruncana linneiana*, *Globotruncana* sp., *Anomalinoideids* spp. and radiolarians (Appendix 10). Based on these fossils Williams and Heryanto (1985) considered that the Payak Formation is of Late Cretaceous (Turonian) age.

One of the samples collected and analysed by Elf Aquitaine Indonesie from the Melawi Basin sequence during exploration for hydrocarbons in West Kalimantan (B. Porthault, pers. comm., 1987) is from the Payak Formation. It produced an assemblage dominated by angiosperm pollen. This assemblage includes *Alnipollenites* sp., *Lanagianopollis microrugulatus*, *Alnipollenites verus*, *Quercoidites* sp., *Dicolpopollis elegans*, *Gothanipollis bassensis*, *Ignanura* sp., *Victoria* sp., *Striatocolpites conspicuus*, *Caryapollenites* sp., *Gemmatriocolpites pergemmatus*, *Florschuetzia trilobata*, *Proxapertites operculatus*, *Ctenolophonidites costatus*, *Sapotaceae*, *Myrtaecidites* sp. and *Ulmipollenites* sp. (Appendix 10). This assemblage indicates a Late Eocene age. Reworked Paleocene and Cretaceous fossils are also present in this assemblage.

The palynological assemblage yielding the Late Eocene age came from the base of the Payak Formation, whereas the rest of the formation has yielded only non-diagnostic foraminifers and
molluscs. Due to its great thickness it is possible that the age of the Payak Formation may range up into the Oligocene. The presence of abundant small bivalves indicates shallow marine depositional environment.

The Payak Formation is conformably overlain by the Tebidah Formation. Twenty samples from the Tebidah Formation were analysed for microfauna: ten of them were barren; nine contained only gastropods and bivalves (with ostracods in one of the samples); and one sample contained indeterminate planktonic foraminifers including one possible Hedbergella trocoidea (A.R. Lloyd, pers. comm., 1984).

Gastropods and bivalves present in the Tebidah Formation have not been indentified, but Icke and Martin (1905) and Van Emmichoven (1939) described the fossils as a lagoonal and brackish water Early Tertiary fauna. No samples from the Tebidah Formation were taken for palynological analyses by Elf Aquitaine Indonesie during petroleum exploration in the Melawi Basin. Consequently the Tebidah Formation has no accurate determination.

The Tebidah Formation is overlain by the Sekayam Sandstone in the west, and by the Alat Sandstone in the east. No fossils have been found in either sandstone, but based on their stratigraphic position above Late Eocene rocks, they are probably of Oligocene age.

2.4.2 Ketungau Basin

Deposition in the Ketungau Basin began with the Kantu Formation. Eight samples were taken from this formation for
microfaunal analysis; three of them were barren. The rest of the samples mostly contain molluscs (mainly bivalves and some gastropods). The bivalves include forms which appear to be oysters. Echinoid spines and ostracods are also found in this formation. The age is indeterminate but beds rich in oysters are known from the Late Eocene of southwest Java. This is not a firm basis for correlation but could indicate a similar age (A.R. Lloyd, pers. comm., 1984). The appearance of oysters indicates a shoreline (marine) or estuarine environment.

Van Emmichoven and ter Bruggen (1935) identified specimens of *Cyrena* (*Batissa*) *borneensis* Botger and *C.* (*Corbicula*) *pengaromensis* Botger which indicated an Eocene age. In the laterally equivalent Silantek Formation in Sarawak, Tan (1979) found *Batissa subtrigonalis* (Krause), *Corbula* (sensu lato), *Vivaparus* or *Paludomus*, *Naticacea* and *Turritellidae*. Based on these fossils he suggested that the Silantek Formation is Late Eocene and may range up into the Oligocene or even Miocene.

The Kantu Formation is conformably overlain by the unfossiliferous Tutoop Sandstone that is intruded by the Late Oligocene to Early Miocene Sintang Intrusives. The Tutoop Sandstone is, therefore, possibly Late Eocene or Oligocene in age. It is correlated with the Alat Sandstone which lies on the Late Eocene-Oligocene Tebidah Formation. Based on these stratigraphic relationships the age of the Tutoop Sandstone is most probably Oligocene.

The Tutoop Sandstone is conformably covered by the Ketungau Formation. Gastropods and bivalves found in the Ketungau Formation have not so far been identified. Van Emmichoven
(1939) also found echinoids and foraminifers which are not age diagnostic. The Ketungau Formation is also intruded by Late Oligocene to Early-Miocene Sintang Intrusives and is probably of Oligocene age.

2.5 STRUCTURAL GEOLOGY

The geology and gravity maps of West Kalimantan (Figs 1.3 and 2.27) indicate the east-west oriented linear nature of structural trends. Williams et al. (1988) divided the west Kalimantan region into three geological domains based on the differences in structural and geophysical pattern, and the distinctive geological characteristics. These domains are the Schwaner Mountains, Northwest Kalimantan Domain, and Boyan and Lubok Antu Melanges with their associated accretionary deposits (Fig. 1.4).

Several major faults along two dominant trends are present in the study area. Faults trending west-northwest are generally parallel to formation boundaries, whereas the set trending east-northeast cross-cut these boundaries (see Geological Map, Appendix 1). The two major faults controlling the structural development of the region are the fault separating the Ingar and Selangkai Formations, and the fault complex on the southern side of the Ketungau Basin (Appendix 1). Activity on these faults probably produced the uplift of the Selangkai Formation prior to deposition of the Suwang Group. The fault between the Selangkai Formation and Boyan Melange is thought to be a high-angle reverse fault. This fault does not affect the Alat Sandstone and thus has been
inactive since at least the Oligocene. The east-northeast trending faults offset the southern fault but appear to merge with the Ketungau graben faults.

More detail, based on the differences in geophysical pattern, structural and geological characteristics and histories, has enabled the study area to be divided into seven structural units (Fig. 2.28) as follows:

2.5.1 Northwestern Kalimantan Domain

The Northwest Kalimantan Domain (Williams et al., 1988) in the study area is composed of Permian granitoid and metamorphic rocks, which are covered unconformably by Late Triassic to Early Jurassic continental to shallow marine sedimentary rocks (Tan, 1986) which, in turn, are unconformably overlain by Jurassic to Cretaceous strata.

The disconformity between the Late Triassic and Jurassic sedimentary sequences (Fig. 2.2) is based on the existence of an hiatus between those units (Tan, 1986). However, the Late Jurassic to Early Cretaceous unconformity and deformation event are thought to result from the Yenshanian Orogeny (Williams et al., 1988; Fig. 1.4).

2.5.2 Schwaner Zone

The Schwaner Zone consists of tonalite, granodiorite and granite with minor low grade metamorphic rocks. According to Williams and Harahap (1987), chemical analyses of typical rocks from the Schwaner Mountains indicate the I-type
calc-alkaline nature of the suite. This unit acts as the basement of the Melawi Basin.

2.5.3 Submarine Cretaceous Zone

The zone extends across the study area in a belt up to 45 km wide, and includes the Selangkai Formation. The general trend of folding in the zone is west-northwest. The eastern part of the Selangkai Formation shows angular folds causing local overturning of beds. North of the Silat Syncline, folding is characterized by multiple cleavage development and the formation of thrust faults.

2.5.4 Boyan Melange Zone

The dominant pervasive shearing in the melange is generally parallel to the boundary faults. The fault separating the melange from the Selangkai Formation is exposed in the Siberuang River as a fault zone which is strongly sheared, and layers of sandstone boudins and volcanic clasts have been incorporated into the shear zone. The presence of ultramafic detritus in sediment from the Ketungau Basin and the abundance of chert and minor mafic grains in the Alat Sandstone suggest that the Melange Zone formed a structural high from Eocene to Miocene times. It may have also formed the northern barrier of the Melawi Basin during the Early Tertiary. The melange zone includes the region referred to by previous workers as the Semitau High.
2.5.5 Melawi Basin

The Melawi Basin was defined as a structural basin by Williams and Heryanto (1986). This structural basin can be divided into three subunits as follows: Pre-Silat Fold Belt, Silat Fold Belt, and Post-Silat Fold Belt.

2.5.5a PRE-SILAT FOLD BELT

This structural subunit includes the gently dipping Ingar Formation consisting of slightly fractured rocks that are locally deformed. Limestone and sandstone blocks, formed by boudinage and fracturing, are up to 1 m in length and often show a good cleavage. The blocks show a weak preferred orientation and are widely dispersed.

2.5.5b SILAT FOLD BELT

This structural subunit forms the narrow fold belt on the southeastern margin of the Semitau High (Selangkai Formation and Boyan Melange). It contains the Dangkan Sandstone and Silat Shale (Williams and Heryanto, 1986) of the Suwang Group. It is folded into a tight east-plunging asymmetrical syncline with a steep northern limb and a gentle southern limb; the limbs are in places overturned. The syncline at Mt Sebuluh appears to be the remnant of a related structure. The two synclines form a southward verging pair of folds with the anticlinal region erosionally removed. Within the fold belt the Silat Shale is strongly folded producing large-scale chevron folds only visible when thin siltstone beds are present. Williams et al. (1984) suggested that the nature of
the folding is controlled by the presence of a thrust fault at the depth. Slickensiding and internal shear planes are also common.

2.5.5c POST-SILAT FOLD BELT

Rocks above the Silat Fold Belt are separated from the underlying Ingar Formation and Suwang Group by an unconformity. Rocks in this structural subunit represent a different sedimentary facies and are always less deformed than the underlying formations.

The northern boundary of the Melawi Group is exposed as the south dipping limb of the large synclinal structure. In the Kapuas and Melawi Rivers the Melawi Group is present in the core of the syncline.

The Melawi Group was folded into a gentle syncline before the Sekayam and Alat Sandstones were deposited. The angular discordance between the Sekayam Sandstone and the Tebidah Formation decreases to zero in the south, where the formations are paraconformable. In the core of the syncline the Tebidah Formation is very gently folded and the junction with the Alat Sandstone is probably paraconformable.

2.5.6 Lubok Antu Melange Zone

The Lubok Antu Melange extends into Kalimantan from Sarawak. The Danau Mafic Complex, which is poorly exposed in the Lake District Depression, is probably a component of this melange. The rocks are strongly fractured in the Lake
District, but no details about the structural evolution of this zone are available in the study area.

2.5.7 Ketungau Basin

The Ketungau Basin is faulted against the Lubok Antu Melange in the north, and forms a large synclinal structure with a width of up to 50 km. Dips on the northern and southern margins are generally about 30°. The southern boundary is also affected by faulting, but onlap of the lowest formation onto the Boyan Melange is also apparent. The thick Tutoop Sandstone defines the structure well in the Ketungau region (western part), but in the eastern part of the basin the structure is not well defined and cannot be related to the Ketungau Basin Syncline. The structure in the eastern part of the basin consists of a north-dipping sequence (with dips up to 55°), which also appears to onlap the Boyan Melange.
CHAPTER 3
SEDIMENTATION

3.1 SEDIMENTARY STRUCTURES

The Melawi and Ketungau Basin sequences consist of alternating shallow marine to terrestrial deposits that indicate tectonic instability during the depositional history of the two basins. Tectonic influence during deposition of the Melawi Basin rocks is also reflected by the presence of three unconformities within the sequence, whereas the slightly younger Ketungau Basin sequence is conformable.

Sedimentary structures preserved within the Melawi and Ketungau Basin sequences can be divided into two groups; those produced by physical mechanics and those produced by biologic activity. Physical sedimentary structures include graded bedding, flute and load casts, tool marks, small- to medium-scale cross-bedding, ripple marks and slump structures. Burrows and bioturbated sequences provide clear evidence of biogenic activity in some of the units.

Primary sedimentary structures are formed at the time of deposition or shortly thereafter. The character of these structures depends mainly on current velocity, water depth, grain size and sedimentation rate (Lindholm, 1987).

An association of particular sedimentary structures within a formation reflects the depositional environment of that unit. A detailed environmental analysis of each formation is given in Chapter 7.
3.1.1 **Hydrodynamic conditions from sedimentary structures**

Hydrodynamic conditions can be inferred from a variety of syndepositional sedimentary structures. The most relevant sedimentary structures for such determinations in the Melawi and Ketungau Basin sequences are cross-beds which can be used to infer possible flow depths and velocities.

From experimental and field measurements Allen (1970a) formulated a relationship between dune size and flow depths for dunes more than 10 cm high. The derived empirical formula was $H = 0.086D^{1.19}$, where $H$ is the dune amplitude and $D$ is the flow depth. Thus the amplitude or the height of the preserved cross-strata set can be related to a minimum depth of flow required for the formation of the set. The actual depth could be greater than the calculated value, since erosion of the top of the cross-stratified set is common.

An alternative method of estimating flow depths is to use the approximate ratio of 1:6 for dune height to flow depth (Collinson and Thompson, 1982), although this ratio should be treated with caution as it is based on a two-dimensional model. Collinson and Thompson suggested that a ratio nearer 1:2 may be more appropriate for relating dune height to flow depth.

The formula for the dimensionless Froude number $(Fr)$ is $Fr = u / \sqrt{gD}$, where $u$ is the velocity, $g$ is the acceleration due to gravity and $D$ is the flow depth (Allen, 1970b). Thus if a flow depth has been calculated and the Froude number is estimated, based on the form of the cross-bedding, then an
estimate can be obtained for the flow velocity at the time of
dune formation. Diagrams in Leeder (1982) and Blatt et al.
(1980) indicate the relationship between velocity and flow
depth (Figs 3.13, 3.14 and 3.15) and show that dunes of
medium-grained sandstone form in a regime where Froude
numbers are between 0.3 and 0.8. The most appropriate Froude
number to use in any given situation can be estimated from
the angle and basal contact of the foreset laminae. Cross-beds with angle of repose foreset laminae having an
angular basal contact are typical of the lower part of this
Froude number range whereas cross-beds with low angle
asymptotic foreset laminae would form in the upper part of
this range.

The diagram in Blatt et al. (1980), which shows the
relation between mean size (mm) and mean velocity (cm/s;
Fig. 3.16), can also be used to calculate the minimum flow
velocity required to transport material with a given mean
grain size.

Blatt et al. (1980) also presented a bedform phase diagram
which shows the relation between grain size, stream power and
bedforms (Fig. 3.17), while Leeder (1982) presented a similar
diagram which shows the relationship between grain size, bed
shear stress and bedforms (Fig. 3.18). These diagrams can be
used to infer the velocity of the flow from the bedform and
the grain size of the sediment.

Allen (1970a) also presented a formula for relating dune
wavelength (\( T \)) and flow depth (D), i.e. \( T = 1.16D^{1.55} \) and
the equivalent relationship was shown graphically by Leeder
(1982; Fig. 3.19). Since the estimate of flow depth used in this analysis is based on preserved cross-bed height these relationships really relate cross-bed height to dune wavelength. However, an estimate of the size of the dunes is useful in terms of environmental considerations.

3.1.2 Melawi Basin

3.1.2a INGAR FORMATION

The predominantly fine-grained Ingar Formation contains rare slump folds, some pull-apart structures and limestone blocks (slide deposits). Parallel lamination and load-casts are common within sandstone beds. Bioturbation structures are found within siltstone and mudstone beds.

Slump folds are found within the siltstone to very fine-grained sandstone in the upper part of the Ingar Formation (Fig. 3.1). The size of these slump folds averages 1 m high and 2 m in a down-slump direction. Conybeare and Crook (1968) suggested that two distinct types of slump structures can be recognized. The first type of slump structure shows crenulation of the inclined laminae, and is probably attributable to simple gravity slumping. The second type is the slump structure which has the upper part of the laminae folded forward to give a recumbent syncline with the axis normal to the current and the hinge directed up-current. Allen and Bank (1972) identified three types of folded deformation of cross-bedding (Fig. 3.2) as follows: simple recumbent folds, a series of folds with or without overturning, and a combination of faulting, folding and local
destruction of bedding. The first two are common in water-laid sediments while third is common in aeolian dune sediments.

The slump structures within the Ingar Formation can be included in the first type of slump structure defined by Conybeare and Crook (1968), whereas using the classification of Allen and Bank (1972) these sedimentary structures are included as slumps consisting of a series of folds with or without overturning.

Pull-apart structures (Fig. 3.3) are also present in the upper part of the Ingar Formation. The structures involve strata up to about 25 cm thick. Conybeare and Crook (1968) defined pull-apart structures as various forms of disrupted bedding which are commonly associated with flame structures. Flame structures have not been recognized in the Ingar Formation pull-apart structures which consist of well separated sandstone blocks with abrupt subvertical terminations. These structures are attributed to disruption of semi-lithified sandstone beds along incipient joints, probably due to slow gravity sliding on a gently sloping substrate. Sliding may have been initiated by sedimentary loading or by seismic or tectonic movement.

Parallel lamination is common within the fine-grained sandstone and siltstone beds. Reineck and Singh (1980) defined that a lamina is produced as a result of some minor fluctuation in depositional energy under rather constant physical conditions. The dominance of very fine-grained strata indicates that most deposition occurred from
suspension and the laminae, especially the fine sandy laminae, would represent periodic increases in bottom traction current activity.

Load casts occur at the bottom of some sandstone beds in the Ingar Formation and suggest that these thicker sandstone beds were rapidly deposited over weakly consolidated mud. According to Conybeare and Crook (1968), the sand has a lower porosity and is more dense than the mud; consequently the sand tends to sink into the mud, which is displaced in a thixotropic manner. Collinson and Thompson (1982) suggested that load casts form as a result of gravity settling of unstable sediments. The instability is caused by the high porosity and lack of compaction in the silty beds and the greater density of the rapidly deposited overlying sand. Reineck and Singh (1980) suggested that in some cases the sand layer will be broken up into several pillow-shaped or ellipsoidal masses, their size ranging from a few centimetres to several metres. They may be slightly connected, or completely isolated floating freely in a muddy matrix (Fig. 3.5). Such ball-and-pillow structures (Fig. 3.4) are present within mudstone beds in the Ingar Formation, and their size ranges from 10 cm to 30 cm.

Bioturbation structures found within the Ingar Formation appear as subvertical tubes with a 5 mm to 10 mm diameter and up to 3 cm to 5 cm long. Reineck and Singh (1980) referred to Schafer (1956, 1972) and divided bioturbation structures into two major groups. Firstly Fossitextura devormativa (deformative bioturbation structures), which appear without
definite form as mottled structures or irregular flecks of different grain size or colour. Secondly Fossitextura figurativa (figurative bioturbation structures), which possess definite recognizable form, such as burrows. Most bioturbation structures which occur in the Ingar Formation are included in the Fossitextura devormativa category but the subvertical tubular burrows probably represent Arenicolites.

In the upper part of the Ingar Formation at the junction of the Ingar and Kayan Rivers (Fig.3.8) the sequence of the sedimentary structures is as follows: at location 87RH70 (the lowest part of the sequence in that area) the section is dominated by mudstone, especially calcareous grey mudstone, containing some burrows which indicate that it was deposited in a calm marine environment. The intercalated fine- and medium-grained sandstone generally has a sharp basal contact suggesting that these beds represent storm redeposited sand. The grain size and thickness of the sandstone intercalations depend on the storm intensity and the distance of the storm and shore from the depositional site. The thick beds and coarse grain size indicated that deposition probably occurred close to the source area.

Higher in the sequence (at location 87RH71) large slump structures (Fig. 3.1) within very fine-grained sandstone are up to 1 m high and 2 m wide. These slump structures indicate a low to moderate depositional slope where simple gravity slumping (Conybeare and Crook, 1968) produced a series of folds without overturning (classification of Allen and Bank, 1972; Fig. 3.2). This gravity slumping may have
been caused by sediment overloading but more probably it was triggered by an earthquake related to early stages of uplift of the basement to the north of the basin (Semitau High). Gravity induced features were also recorded higher in the sequence at locations 87RH68 and 87RH78. The presence of the pull-apart structures (Fig. 3.3) and limestone blocks (e.g. Fig 3.9) within the mudstone matrix in the lower part of these sections indicates that synsedimentational gravity redeposition was important in this area throughout the period of Ingar deposition.

Towards the top of the formation (e.g. locations 87RH68 and 87RH78) sharp based medium-grained sandstone beds reappear in the sparsely bioturbated mudstone sequence and increase in abundance upwards. These sandstone beds represent a second occurrence of storm redeposited material and indicate shallower water, higher energy influences during the latter stages of deposition in this area.

3.1.2b DANGKAN SANDSTONE

Cross-bedding is an important sedimentary structure in the Dangkan Sandstone. Lindholm (1987) classified cross-bedding on the basis of the shape and attitude of foreset beds, the shape of set boundaries, and the thickness of the sets. Foresets are described as planar or tangential (Fig. 3.11). Planar foresets are nearly flat and intersect the lower set boundary at a relatively high angle, whereas tangential foresets are curved, concave-up and intersect the lower set boundary at a low angle. On the basis of the configuration
of the set boundaries Lindholm described the cross-beds as tabular or trough-shaped (Fig. 3.12). Tabular cross-bedding is characteristically either planar or parallel, whereas trough-shape laminae have strongly curved concave-up lower set boundaries. Based on the thickness of the sets he described them as large scale if the set thickness exceeds 5 cm and small scale if the sets are less than 5 cm thick. Conybeare and Crook (1968) classified cross-bed set sizes into small scale (up to 5 cm), medium scale (5cm - 2m), large scale (2m - 8m) and very large scale (more than 8 m).

The thickness of cross-bedded sets in the Dangkan Sandstone ranges from 20 cm to 30 cm and they represent medium scale cross-beds using the classification of Conybeare and Crook (1968). Some of the cross-beds have moderately to strongly asymptotic foreset laminae indicating deposition under relatively high flow conditions in the lower flow regime. Most cross-beds are characterized by high angle foreset laminae with dips ranging from 20° to 30° (classification of Conybeare and Crook, 1968), as shown in Figure 3.10. Some of the smaller cross-bed sets composed of finer grained sandstone also have planar foreset laminae and indicate slower or deeper flows where the sediment is subjected to less bed shear stress. Graded laminae are very common in the upper and middle parts of the Dangkan Sandstone in both the planar and trough cross-beds (Appendix 2). The thickness of graded laminae ranges from 10 mm to 30 mm.

Using the hydrodynamic methods discussed in Section 3.1.1, the minimum depth of flow that formed the medium scale
cross-stratified sets in the Dangkan Sandstone was probably between 1 m and 2.5 m for both the trough and planar cross-beds, with a maximum calculated flow depth of 4 m (Table 3.1). Based on these data, the most probable flow depth was less than 1.75 m. Using this average flow depth and a Froude number of 0.5 the flow velocity for deposition of the cross-bedded sandstone units was about 2 m/s (ranging between 0.6 m/s and 5 m/s using extreme depth values; Table 3.1). The wavelengths of the dunes which formed the cross-beds in the Dangkan Sandstone varied between 0.3 m and 36.2 m (Table 3.1). Using the mean flow depth of 1.75 m, the minimum and maximum mean dune wavelengths are 7.0 m and 15.7 m (Leeder, 1982).

A few apparently massive beds in the Dangkan Sandstone show grading (or rarely reverse grading) over intervals of 10 cm to 30 cm. These beds contain little matrix and the grading is restricted to the coarse fraction of the grain size — commonly from granule or pebbly sandstone to medium- or coarse-grained sandstone. These beds probably represent high energy plane bedded deposits where the individual laminae cannot be distinguished because of the generally coarse grain size.

Parallel bedding found within some of the finer grained sandstone beds shows parting lineation (Fig. 3.20). This indicates periods of high flow velocities and upper flow regime conditions in shallow water depths (Collinson and Thompson, 1982).
Mudstone rip-up clasts and basal scours are found at the base of some sandstone beds. They are produced as a result of erosion of a sediment surface by the current flowing over it (Reineck and Singh, 1980) or by the lateral erosion of previously deposited sequences.

The appearance of small-scale ripple cross-strata within the finer grained sandstone beds indicate deposition by currents which were strong enough to exceed the initial erosion velocity for the sand but not strong enough or deep enough to form dunes (Collinson and Thompson, 1982). These ripples respond only slightly to flow depth, but their size is directly related to flow velocity (Blatt et al., 1980; Leeder, 1982). The Froude number needed for their formation ranges from 0.1 to 0.4, less than that for dunes (Leeder, 1982). Assuming a flow depth of 1.75 m (probably excessive), the range of flow velocities for ripples formation is from 0.4 m/s to 1.65 m/s.

The stratigraphic column of the Dangkan Sandstone at the type section (Fig. 3.21; Appendix 2) shows that the lower part of the formation consists of massive to thickly bedded coarse-grained to pebbly sandstone and conglomerate. This part of the unit was probably deposited rapidly with the thick conglomerate beds representing gravel bars. The conglomerates are generally covered by well bedded medium- to coarse-grained sandstone which may be partly conglomeratic or pebbly. The central part of the formation is slightly finer grained and consists of well developed fining upwards cycles 4 m to 6 m thick. Higher in the formation a second sequence
of conglomeratic and coarse-grained pebbly sandstone was also followed by a sequence of medium- to coarse-grained sandstone that shows fining upward cycles near the top of the formation.

A typical section of the medium- to coarse-grained sandstone sequence from the central part of the Dangkan Sandstone is shown at locations 106, 107 and 108 (Fig. 3.21). The section starts with a conglomeratic coarse-grained sandstone bed (60 cm thick) that overlies a sharp erosional surface. This bed is followed by 40 cm of structureless coarse-grained sandstone covered by 350 cm of massive fine- to medium-grained sandstone showing normal grading in the lower part and trough cross-bedding (30 cm thick sets; e.g. Fig. 3.10) in the upper part. The latter probably represents channel-base trough cross-stratification. The trough cross-beds are overlain by plane bedded sandstone interspersed with planar cross-bed sets (e.g. Fig. 3.20) which represent fluctuating flow velocities or depths and the migration of small transverse tabular bars. The sequence from conglomeratic coarse-grained sandstone up to planar cross-beds shows a fining upward character which suggests that this sequence represents a channel deposit with a minimum thickness of about 5 m.

This channel deposit was followed by a multistory sandstone body which probably consists of a second channel sand deposit (about 3 m thick) erosionally truncated and overlain by a third fining upward channel sequence (5.3 m thick). The second channel deposit comprises a series of
medium- to coarse-grained sandstone beds which show slight fining upward with individual bed thickness ranging from 35 cm to 60 cm. The third set of channel deposits consists of a series of medium- to coarse-grained sandstone beds overlying a thin pebbly lag deposit. The sequence becomes finer upward with the bedding thickness ranging from 30 cm to 50 cm. The upper part of the sequence contains parallel laminated beds, planar cross-beds and ripple marks.

3.1.2c SILAT SHALE

Ripple marks, small scale festoon cross-bedding, planar cross-bedding, graded laminae, slumped lamination, tool marks and load casts are present within sandstone interbedded in the Silat Shale.

Symmetrical straight crested ripple marks and some ripple cross-laminae are found in the very fine-grained sandstone beds. The ripple heights are approximately 20 mm with a wavelength of approximately 50 mm. The symmetrical nature of the ripples suggests wave action during deposition in a shallow standing body of water (Collinson and Thompson, 1982).

Small-scale festoon cross-bedding (Conybeare and Crook, 1968) is also found within the very fine-grained sandstone beds. The set thickness ranges from 3 cm to 5 cm and the troughs are generally 10 cm to 20 cm wide. In some places it is associated with slumped or convolute lamination.

Medium scale planar cross-bedding (Conybeare and Crook, 1968) is a minor component usually found within the very
fine- to fine-grained sandstone beds. The average set thickness is about 10 cm and the maximum set thickness is 15 cm, with the angle of dip ranging from 20° to 30°.

The cross-bed thickness is near the minimum for calculating flow depths using the formula of Allen (1970a). Calculated flow depths range up to 1.6 m (Table 3.1) but most estimates are less than 1 m. The corresponding flow velocities for deposition of the cross-beds range from 0.3 m to 3.2 m/s (Table 3.1), with the most probable flow velocities being less than 1.2 m/s. Dune wavelengths for generating the cross-beds vary between 0.2 m and 14.3 m.

Graded laminae are present within some very fine-grained sandstone and siltstone beds (Fig. 3.22). The set thickness ranges from 1 cm to 5 cm. This sedimentary structure is commonly associated with load casts, parallel lamination and convolute lamination (Fig. 3.22). The genesis of convolute lamination in normally graded laminated beds occurs under the driving force of gravity. It is produced after liquefaction and involves passive fluid movement while the deposit is in a fluidized state (Allen, 1977). The disruption of the grain fabric may occur as a result of seismic shock, an increase in pore-water pressure as a result of movement of formation water, or pressure changes due to the passage of internal or surface waves. Convolute lamination is a typical feature in subaqueous slope deposits and in turbidite sequences. Middelton and Hampton (1973) considered that partial liquefaction of very fine- to fine-grained sand was the main cause of convolute bedding in turbidite sequences, and
liquefaction can take place at several intervals during the deposition of the bed.

Tool marks (Fig. 3.23) and flute marks are also found at the base of some sandstone beds in the Silat Shale. Reineck and Singh (1980) divided tool marks into three types as follows: stationary tool marks, obstacle marks and moving tool marks. The tool marks present in the Silat Shale mainly consist of obstacle marks. From their experiments Dzulynski and Walton (1965) concluded that flute marks are produced on a sediment bed by the turbulent action of either sediment-laden currents (turbidity currents) or sediment-free currents. They also suggested that in all cases the flute marks are formed in the strongly turbulent zone near the base of the flow. The presence of both tool and flute marks in the Silat Shale suggests that at least some of the sandstone beds were deposited by turbulent mass flow mechanisms. According to Middleton and Hampton (1976) there are four types of sediment gravity flow (Fig. 3.24): turbidity current flow, liquefied sediment flow, grain flow and debris flow. The sedimentary structures which represent each type of sediment gravity flow are illustrated in Figure 3.25.

Flame structures are also present within some very fine- to fine-grained laminated sandstone beds in the Silat Shale. This sedimentary structure is produced by water escape associated with rapid sediment loading. Lowe (1975) distinguished three main processes for forming flame structures. Firstly, seepage is the slow upward movement of pore water without disturbing the sediment grains. Secondly,
collapse of a loosely packed framework of sediment grains causes liquefaction and the grains become suspended in their own connate fluid. The fluid is displaced upward, and a more tightly packed grain-supported fabric is produced. Finally, fluidization of a sediment can occur when a moving pore fluid exerts an upward drag force on the sediment grains, lifting the grains and destroying the grain fabric.

Evidence from the association of sedimentary structures both within and beneath the graded sandstone beds in the Silat Shale strongly support a low concentration turbidity current depositional mechanism for these units whereas the cross-stratified beds represent traction current deposits.

The vertical sequence of sedimentary structures in the type section of the Silat Shale is shown in Fig. 3.26 and in Appendix 3. The lower part of the formation consists predominantly of massive shale although a few thin scattered lenses and beds of fine-grained sandstone are also present. Parallel lamination, including carbonaceous laminae, is the only sedimentary structure recorded in the lower part of the formation within the dark grey to blackish mudstone.

Most of the distinctive sedimentary structures in the Silat Shale were found within the upper part of the formation.

In the section at location 87RH100 (Fig. 3.26) the sedimentary structures include parallel lamination together with some ripple, slump (e.g. Fig. 2.12) and convolute lamination (Fig. 3.22). This association of sedimentary structures occurs in a very fine-grained sandstone sequence
with a total thickness of 18.5 m. These sedimentary structures suggest that deposition occurred on a gently sloping substrate at some distance from the source of sediment supply — possibly on a distal delta margin or in a shallow trough area where the substrate was not influenced by waves or nearshore currents. The overlying 1 m thick medium-grained sandstone bed shows a sharp basal contact contact with some tool marks (Fig. 3.23) indicating that scouring occurred at the time of deposition. This sandstone bed probably represents a mass flow redeposited sand, and indicates a period of high energy in the source region that may be related to storm waves or a river flood. Medium-grained sandstone beds become more common in the upper part of this section where they range in thickness from 40 cm up to 80 cm, and contain sedimentary structures such as ripple marks, slump structures and parallel bedding. The appearance of more numerous sandstone beds indicates that influxes of high energy conditions increased with time, which in turn suggests that the environment of deposition became shallower and was closer to the source of high energy events. A coarsening upward sequence such as this typically occurs in prograding deltaic environments and indicates a relative decrease in sea level, or regression.

The measured section at location 87RH102 (Fig. 3.26) is similar in many respects to the section at location 87RH100. The lower part of the section consists predominantly of laminated shale with a few very fine- to fine-grained sandstone beds containing sedimentary structures such as
parallel, convolute and cross-laminae. These structures indicate periodic traction and mass flow movement of material into an otherwise low energy depositional environment. A couple of thicker massive to weakly graded medium-grained sandstone beds in the upper part of the shale suggest a more proximal character and the depositional environment may have become shallower. A major break at the top of the shale is recorded by a 90 cm thick coarse-grained conglomeratic sandstone bed overlying an erosional surface. The conglomeratic bed is overlain by a fining upward sequence which includes 2 m of well bedded medium-grained sandstone (with bedding thickness ranging from 15 cm to 30 cm) and about 1 m of thinly bedded fine-grained sandstone. The conglomerate is very similar in form and character to conglomeratic beds described from the Cardium Formation in western Canada by Walker and Eyles (1988, 1991). The latter beds were attributed to subaerial exposure and fluvial transport of the coarse clastic detritus during a regressive interval, which was drowned and reworked during the succeeding transgression. The upper part of the section consists of a 1 m thick coarse-grained massive sandstone bed capped by a 25 cm thick tuffaceous bed and 2 m of well bedded (10-20 cm thick) fine- to medium-grained sandstone. This part of the section indicates a general decrease in energy that may be associated with the transgression.
3.1.2d SEPAUK SANDSTONE

Only a few sedimentary structures have been observed within the Sepauk Sandstone due to limited time for additional field work in this area. Medium scale cross-beds (Conybeare and Crook, 1968) are common in the Sepauk Sandstone. Trough cross-beds, with an average set thickness of about 0.24 m, and planar cross-beds, with an average set thickness of about 0.25 m, have dip angles of about 30°. The maximum cross-bed set thickness is 0.50 m (Table 3.1).

The flow depths calculated for cross-bed deposition range from 0.5 m to 4.4 m (Table 3.1), with the most probable depth ranging from 1 m to 2.5 m. The flow velocity for deposition of the cross-beds ranges from 0.3 m/s to 5.25 m/s (Table 3.1). The most probable flow velocities were between 0.5 m/s and 1 m/s. Dune wavelengths associated with these cross-beds probably ranged from 5 m to 15 m with extreme values ranging between 0.4 m and 39.5 m (Table 3.1).

The lower part of the Sepauk Sandstone is characterized by the presence of poorly defined fining upwards sequences starting with coarse-grained, sometimes pebbly sandstone, passing upward into finer grained flat and cross-bedded sandstone. Other sedimentary structures recorded from the Sepauk Sandstone include scour surfaces which are common between the sandstone beds, especially at the base of fining upward cycles. In such areas erosion has cut down into the finer grained underlying sandstone beds of the preceding sequence.
3.1.2e PAYAK FORMATION

Sedimentary structures which are present within the Payak Formation include parallel lamination, cross-bedding and ripple marks. Linear groove casts are present on the base of some beds and the upper surfaces of beds are commonly rippled.

Cross-beds which occur within the Payak Formation consist of trough and planar cross-beds. The medium scale (Conybeare and Crook, 1968) cross-beds have an average set thickness of about 0.12 m for trough cross-beds and 0.16 m for planar cross-beds, and the maximum set thickness is 0.30 m.

Flow depths for the formation of the medium scale cross-bed sets in the Payak Formation would probably have been within the range from 0.25 m to 1 m, with a maximum indicated depth of almost 3 m (Table 3.1). Calculated flow velocities for the formation of the cross-beds varies from 0.4 m/s to 4.2 m/s (Table 3.1) with most of the data indicating a velocity of less than 0.9 m/s. Corresponding dune wavelengths range from 0.2 m to 25.7 m (Table 3.1).

The Payak Formation contains carbonaceous material, palynomorphs and, in places, abundant small bivalves. The unit is dominated by flat and cross-bedded sandstone but contains some silty intervals and the base of the formation is coarser grained and conglomeratic in many areas (Fig. 3.27).

The section at location 87RH61 is typical of the lower sandstone dominated part of the formation. The sequence consists of thin poorly defined fining upward units starting
with massive medium-grained sandstone (occasionally pebbly) above a scoured erosion surface, followed by a series of well bedded (10-20 cm thick) fine-grained sandstone units. One of these sequences contains a planar cross-bed set 40 cm thick. The four repetitions of fining upward cosets suggests a multistory origin for the sandstone. Since the individual cosets are only about 2 m thick the sequence may represent a succession of channel base deposits or may represent the migration of successive sand bars in either a channel (fluvial or tidal) or a moderate energy nearshore marine environment.

Location 87RH58 is representative of the central portion of the Payak Formation and consists of interbedded sandstone and mudstone (Fig. 3.27). The sequence starts with a thick (9.6 m) uniform well bedded medium-grained sandstone. Individual beds range in thickness from 50 cm to 75 cm and are internally structureless. This sequence is covered by 4.8 m of mudstone with a few intercalations of siltstone. Bed thickness ranges from 5 cm to 10 cm with a few beds showing internal lamination. This sequence is overlain by 3.6 m of fine-grained sandstone which contains cross-laminae in the lower part and carbonaceous matter in the upper part. The top of this section consists of 1.5 m of coarse-grained sandstone which is also in the form of internally structureless beds. The uniform massive sandstone beds forming the upper and lower parts of the sequence may represent deposition in a nearshore marine environment. The mudstone sequence lacks bioturbation and probably represents
a low energy oxygen depleted environment — possibly lagoonal. The common occurrence of organic detritus in the fine-grained sandstone suggests a nearby fluvial source supplying land plant material.

### 3.1.2f TEBIDAH FORMATION

Medium scale planar tabular and trough cross-beds are characteristic features in the sandstone beds in the Tebidah Formation. Parallel laminae, ripple cross-beds and symmetrical ripple marks are also present within the very fine-grained sandstone beds.

Medium scale planar and trough cross-beds (Conybeare and Crook, 1968) are common in the lower part of the Tebidah Formation; the average set thickness is 0.17 m for trough cross-beds and 0.23 for planar cross-beds. However, in the upper part of the Tebidah Formation, the average cross-bedded set thickness is greater than that in the lower part of the formation with a maximum set thickness of 2.00 m (Fig. 3.28). The foreset laminae show moderate to high dip angles.

The minimum flow depth (using the formula from Allen, 1970a) was 1.8 m and 2.3 m for producing the trough and planar cross-beds, with a maximum depth of 14 m (Table 3.1). The most probable flow depth was less than 4 m. The calculated flow velocity for deposition of the cross-beds varies from 0.4 m/s to 9.4 m/s but was probably less than 1 m/s (Table 3.1). Dune wavelengths corresponding to the cross-bed set sizes in the Tebidah Formation ranged from
0.2 m to 127 m (Table 3.1); i.e. from large ripples to megaripples or elongate low amplitude sand bars.

Slump structures in the sandstone beds in the upper part of the Tebidah Formation (Fig. 3.31) range in thickness from 40 cm to 50 cm, and they are associate with planar and trough cross-beds.

Symmetrical ripple marks are found within the very fine- to fine-grained sandstone beds in the upper and middle parts of the Tebidah Formation. The average ripple height is approximately 30 mm with a wavelength of approximately 90 mm (Fig. 3.29). Small-scale ripple cross-laminae are also found within the same sandstone beds, especially in the upper part of the formation. The set thickness ranges from 3 cm to 5 cm with a dip angle of about 20°. These sedimentary structures are commonly associated with carbonaceous parallel lamination.

Bivalves and gastropods are present in the lower part of the Tebidah Formation (Fig. 3.32). Bioturbation is also present within mudstone and siltstone beds in the upper part of the formation. In addition to indistinct mottled bedding the bioturbation includes figurative bioturbation structures (Reineck and Singh, 1980) such as horizontal and vertical burrows (Fig. 3.30).

The upper part of the formation is mudstone-rich, containing carbonaceous material and coal.

The sequence of the sedimentary structures within the southern limb of the Tebidah Formation is shown in Fig. 3.32. The formation consists predominantly of fine-grained
sandstone and mudstone with only a few thick sandstone dominated sections. One of the thicker sandstone units in the Melawi River at location 87RH17, begins with a 75 cm thick bed of conglomeratic coarse-grained sandstone which gradually fines upward to medium-grained sand at the top. It probably represents a channel lag deposit. The overlying trough cross-bedded sequence has each individual set overlying a scoured erosional base. The lower cross-beds show minor slumping of the foreset laminae (Fig. 3.31). Some mudclasts are found near the basal scour surfaces indicating lateral erosion of earlier fine-grained overbank deposits. The average thickness of the cross-bed sets is about 1 m. The total channel thickness is about 5 m and the channel shows evidence of lateral migration and accretion (Fig. 3.28).

In the Mentatai River at location 87RH05 the section begins with a 290 cm thick, very fine-grained sandstone bed containing common parallel laminae. This sandstone bed was covered by 3 m of massive siltstone, 75 cm of carbonaceous shale and 1.5 m of siltstone. Lower in the sequence at location 87RH08, a 1.5 m thick very fine-grained parallel laminated sandstone bed at the base of the section is capped by a soil horizon and a 25 cm thick coal seam. The latter two features indicate a moderate period of subaerial exposure. The coal is overlain by 1.2 m of carbonaceous mudstone which contains accumulations of brackish water or marine bivalves and gastropods at the base. Siltstone (2.5 m thick) lies above the carbonaceous mudstone and both sections
represent tidal flat deposits associated with the lower reaches of a river floodplain or delta.

On the Timber Road at location 87RH12 the section dominantly consists of very fine-grained sandstone containing parallel laminae, burrows (Fig. 3.30), planar and trough cross-lamination. The latter have an average set thickness of 5 cm to 10 cm. Intercalations of siltstone (5-10 cm thick) are found scattered through the very fine-grained sandstone. The siltstone is parallel laminated and contains scattered small burrows. A few interbeds of medium-grained sandstone (thickness ranging from 20 cm to 25 cm) are present in the predominantly fine-grained unit and they have sharp upper and lower contacts. The coarser sandstone beds are internally massive. The very fine-grained sandstone part of this sequence may represents overbank or floodplain deposits, whereas the interbedded medium-grained sandstone may represent crevasse splay deposits. On the Timber Road at location 87RH12A, vertically above the section at 87RH12, the outcrop consists of massive and minor parallel laminated reddish purple mudstone (Fig. 2.16) which indicates that oxidation played a major influence during its deposition. It probably represents overbank floodplain or ephemeral floodplain lake deposits. Equivalent mudstone in the Mentatai River at location 87RH006 is greenish grey with some parallel lamination (Fig. 2.17) and may represent lacustrine or lagoonal deposits.

The vertical distribution of sedimentary features in the northern limb of the Tebidah Formation is shown in Fig. 3.33.
In this region the upper and lower parts of the Tebidah Formation are predominantly fine-grained whereas the central part of the formation contains coarser sandstone units.

A typical channel deposit from the central part of the formation in the Kayan River is clearly shown in the section at location 87RH39. The base of the channel is represented by an erosional surface. The channel fill consists of medium- to coarse-grained sandstone which becomes finer towards the top of the sequence. Slumped foreset laminae are present in near the base of the sequence and are overlain by large scale trough cross-beds (in sets up to 2 m thick). Evidence of lateral accretion in the channel fill is present near a few preserved channel margins (e.g. Fig. 3.28). The average thickness of individual fining upward channel deposits is about 6 m.

The section at location 87RH30, towards the top of the formation, is representative of the mudstone dominated part of the formation. The section dominantly consists of mudstone with scattered interbeds of fine-grained sandstone in the lower part and medium-grained sandstone (e.g. Fig. 2.15) in the upper part. The thickness of the sandstone interbeds ranges from 75 cm to 120 cm. The very fine-grained sandstone beds contain parallel lamination and planar cross-lamination, with symmetrical ripple marks recorded in one bed (Fig. 3.29). At location 87RH31, the section is also dominated by mudstone and is generally similar to location 87RH30. In this section, the sandstone interbed is fine- to medium-grained and shows no internal structure. It is
overlain by 110 cm of partly bioturbated carbonaceous mudstone containing an accumulation of gastropods and bivalves at the base. The section at 87RH30 probably represents overbank floodplain deposition interspersed with crevasse splay sandstone units whereas the sections at 87RH31 may represent tidal flat deposits with the sandstone interbeds representing fluvial flood sands from an adjacent channel or thin laterally reworked beach sands.

3.1.2g SEKAYAM SANDSTONE

Cross-bedding and parallel bedding are the dominant sedimentary structures found within the Sekayam Sandstone. The cross-beds include medium scale (Conybeare and Crook, 1968) planar cross-beds with an average set thickness of about 0.24 m and dip angles of about 25°.

Flow depths required for the formation of the cross-beds ranged from 0.5 m to 2.4 m, with a probable flow depth of about 1.5 m (Table 3.1). The corresponding flow velocities vary between 0.25 m/s and 4.5 m/s, with the most probable range between 0.5 m/s and 1 m/s (Table 3.1). The wavelength of the dunes responsible for producing the cross-beds varies between 0.37 m and 29.25 m (Table 3.1).

No fossils have been recorded from the Sekayam Sandstone. The sandstone occurs in poorly defined fining upwards sequences with intercalations of reddish oxidized mudstone present throughout this unit.
3.1.2h ALAT SANDSTONE

Sedimentary structures such as fining upward sequences, cross-bedding and parallel bedding are also present in the Alat Sandstone.

Scour surfaces are present between sandstone beds, especially when coarse beds overlie finer grained beds. Coarse lag deposits are present in some scour pockets at the base of fining upwards sequences and are associated with coarse bedload gravel and sand. The lag deposits give way vertically to trough cross-stratified sandstone and constitute the typical repetitious fining upwards sequences constituting the lower and middle parts of the Alat Sandstone.

Parallel laminae are also found in the lower and middle part of the Alat Sandstone. They occur as isolated sedimentary structures in otherwise massive units or they may be associated with planar and trough cross-beds. Parallel laminated beds show alternations of grain size (Fig. 3.35) with set thickness ranging from a few centimetres up to 20 cm. A few plane bedded sandstone units show primary current lineation and were probably deposited by shallow upper flow regime currents.

Normal grading occurs in some massive sandstone beds, especially in the lower parts of fining upwards sequences. The beds have a set thickness ranging between 15 cm and 30 cm and probably represent material rapidly deposited by high velocity traction currents.
Medium scale trough and planar cross-beds (Conybeare and Crook, 1968) are common in lower part of the Alat Sandstone (Fig. 3.34). The average set thickness is 0.18 m for trough cross-beds and 0.32 m for planar cross-beds, with the maximum thickness being 0.50 m. The dip angles range from $20^\circ$ up to $30^\circ$.

The minimum flow depths calculated from the cross-bed thickness range from 0.4 m to 4.4 m (Table 3.1). The most probable average flow depth was between 2 m and 3 m. Flow velocities for the deposition of the cross-beds could have ranged between 0.3 m/s and 5.25 m/s (Table 3.1), with most beds indicating velocities of about 0.5 m/s to 1.3 m/s. Corresponding dune wavelengths for the preserved cross-beds sets range from 0.25 m to 39.5 m.

Small scale trough cross-stratification (Conybeare and Crook, 1968) and parallel lamination are common within the very fine- to fine-grained sandstone beds in upper part of the Alat Sandstone. Cross-beds set thickness ranges from 30 mm up to 50 mm (Fig. 3.37). These structures are associated with ripple marks, climbing ripples and parallel carbonaceous laminae.

The general distribution of sedimentary sequences in the Alat Sandstone is shown in Fig. 3.36. The Alat Sandstone sequence can be subdivided into three major sections based on stratigraphic height. The lower part of the formation consists of coarse-grained sandstone and conglomerate, the middle part consists dominantly of medium-grained sandstone
and the upper part consists of very fine- to fine-grained sandstone.

The lower Alat Sandstone is represented by the section at location 87RH20. In the lower part of the section repetitive 1 m to 2 m thick units show a fining upward from conglomerate to coarse-grained sandstone (Figs 2.19, 2.20 and 3.35). The aggregate thickness of this coarse sequence is 9 m and it probably consists of a multistory stack of channel base deposits. Each unit was probably deposited rapidly by a migrating longitudinal pebbly mid-channel bar. The coarse sequence was followed by deposition of massive coarse-grained sandstone showing trough cross-bedding in the slightly finer upper part (e.g. Fig. 3.34). A second sequence starts with an abrupt erosional base overlain by about 5 m of massive coarse-grained sandstone with some pebbly intervals showing internal erosional features. This sandstone may also represent multistory channel deposits. Above the coarse sand unit the beds are finer grained with cross-bed sets 30 cm to 50 cm thick. The top of the section is represented by internally massive medium-grained sandstone beds 30 cm to 50 cm thick. These upper sandstone beds probably represent the start of another channel deposit.

The central part of the Alat Sandstone is similar to the lower section but the fining upward sequences are more clearly developed, tend to lack multistory channel sand deposits and contain fewer pebbles and gravel lenses.

The upper part of the Alat Sandstone is represented by the section at location 87RH25. This section starts with 2.5 m
of coarse-grained sandstone and conglomeratic sandstone that probably represents the upper part of a channel deposit. It is overlain by 10 m of very fine- to fine-grained sandstone characterized by the presence of parallel laminations and ripple marks (Fig. 3.37). This alternating parallel and ripple bedded sandstone sequence probably accumulated as splay sands from periodic shallow overbank flows in an area close to a major channel. Such thinly bedded fine sequences dominate much of the uppermost part of the formation.

3.1.3 Ketungau Basin

3.1.3a KANTU FORMATION

Parallel laminated, graded and cross-bedded sandstone are prominent features in the Kantu Formation. Cross-lamination, parallel lamination and carbonaceous laminae are also common within interbeds of siltstone and very fine-grained sandstone.

Parallel laminated and graded beds are most common within the sandstone subunits in the upper and lower parts of the Kantu Formation. Parallel laminated beds are composed of laminae with alternating grain sizes in sets from 3 cm up to 5 cm thick. The thickness of graded beds ranges from 10 cm up to 20 cm and they appear either as isolated sedimentary units or in association with cross-beds.

Medium scale planar cross-beds (Conybeare and Crook, 1968) are also common within the sandstone subunits in the upper and lower parts of the Kantu Formation. The average
thickness of sets is about 0.26 m with the maximum thickness being 0.31 m. Flow depths during deposition of the cross-beds were calculated to be in the range from 0.5 m to 3.25 m with the most probable depth being 1 m to 2 m (Table 3.1). Equivalent flow velocities at the time of deposition were probably between 0.3 m/s and 4.3 m/s but most estimates indicate a velocity of less than 1 m/s (Table 3.1). The dune wavelength for the cross-beds in this formation ranges between 0.42 m and 26.5 m.

Cross-lamination and small scale cross-beds (Conybeare and Crook, 1968) are present within the siltstone and very fine-grained sandstone beds in the middle part of the Kantu Formation. The thickness of cross-laminated sets in this interval ranges from 3 cm up to 5 cm.

Parallel lamination is also found within the siltstone and very fine-grained sandstone beds which form a major constituent in the central part of the formation. The laminae are enhanced by differences in colour, and some of the laminae are carbonaceous.

Echinoids, ostracods, bivalves and gastropods are present in the lower part of the formation, especially in Sarawak (Tan, 1979).

The sedimentary sequence in the Kantu Formation is shown in Fig. 3.38. In general, the Kantu Formation in the Kantu-Kantuaping River area consists dominantly of sandstone in the lower part, whereas the remainder of the formation comprises siltstone and mudstone with intercalations of very
fine-grained sandstone and rare beds of medium-grained sandstone.

A typical section from the sandstone dominated lower part of the formation occurs at location 87RH179 (Fig. 3.38). The section consists of medium- to coarse-grained sandstone with bedding thickness ranging from 50 cm to 150 cm. Not many sedimentary structures have been recorded from this section apart from some fining upward intervals. The multistory sandstone beds are separated by subplanar erosional surfaces and may represent superimposed distal low sinuosity channel deposits or possibly moderate energy nearshore marine deposits.

The section at location 87RH160 represents a typical finer grained sequence from the central part of the Kantu Formation. It starts with 2.5 m of internally structureless coarse-grained sandstone which may indicate rapid deposition. This bed is covered by 2 m of very fine- to fine-grained sandstone containing parallel laminations (e.g. Fig. 2.23) that may represent bar top, levee or proximal overbank deposits. Two meters of well bedded medium-grained sandstone (20-30 cm thick) overlie the fine-grained unit with a sharp erosional contact. The next 2.5 m consists of siltstone containing parallel laminae which may represent overbank floodplain deposits. This siltstone bed is covered by 150 cm of well bedded sandstone (individual beds 20-30 cm thick) which, in turn, is overlain by 650 cm of well bedded and laminated (5-10 cm thick) very fine- to fine-grained sandstone. This unit may also represent overbank or
floodplain deposits. The fine-grained unit is truncated by an erosional surface at the base of the next 3 m thick bed of massive coarse-grained sandstone. It probably represents a rapidly deposited channel base deposit that was capped by a further 3.5 m of well bedded (20-30 cm thick) medium-grained channel-fill sandstone.

The lithology and sedimentary variations in the section at location 87RH149 (also from the central Kantu Formation) are generally similar to the preceding section. The lower sequence represents a 10 m thick fining upward unit ranging from coarse- to fine-grained sandstone. The beds are mainly thickly laminated or internally massive. The second fining upward sequence is about 5 m thick consisting of massive coarse- to medium-grained sandstone representing a channel fill deposit that is overlain by a floodplain sequence of laminated mudstone. The top of the section is composed of medium- and coarse-grained sandstone representing possibly two further channel deposits.

3.1.3b TUTOOP SANDSTONE

Cross-bedding is common in the Tutoop Sandstone, with the mean set thickness being 0.36 m and 0.43 m for trough and planar cross-beds respectively. The maximum set thickness is 0.75 m.

Mean flow depths for deposition of the cross-beded sets ranges from 0.72 m to 3.86 m for both the trough and planar forms (Table 3.1). The thickest cross-bed set yields an estimated flow depth of 4.5 m to 6.17 m. Flow velocities
calculated from the above depth estimates, Froude numbers of 0.3 and 0.8, and bedform phase diagrams mainly range from 0.79 m/s to 2.33 m/s with a maximum calculated velocity of 6.22 m/s.

Basal scours, graded beds and parallel laminae (Fig. 3.40) are also present in the Tutoop Sandstone. Basal scour structures resembling asymmetrical troughs (Fig. 3.39), are generally produced on the channel bottom, with their long axes parallel to current direction (Reineck and Singh, 1980). Fining upward channel-fill sequences consist dominantly of sand, and they are commonly conglomeratic near the base. Graded or internally fining upwards beds are minor with average set thickness ranging between 50 cm and 80 cm and a maximum set thickness of up to 2 m.

The vertical distribution of sedimentary sequences within the Tutoop Sandstone are shown in Fig. 3.40. Generally the Tutoop Sandstone in the Segubung River area can be divided into two parts as follows: the lower part consists of coarse-grained and conglomeratic sandstone, and the upper part which consists predominantly of medium-grained sandstone.

A typical part of the lower Tutoop Sandstone is represented by the section at location 87RH229. This section starts with 2.75 m of conglomeratic coarse-grained sandstone, which is internally structureless but shows an upward decrease in grain size. This bed may represent a channel base deposit. This lower sandstone is truncated along a scoured erosional surface by another coarse-grained channel.
base deposit consisting of 1.75 m of conglomeratic coarse-grained sandstone that gradually becomes finer grained upward (e.g. Fig. 3.39). This basal sandstone was followed by 2.25 m of well bedded (50-75 cm thick) coarse-grained sandstone and 2 m of medium-grained sandstone containing trough cross-beds in the lower part and parallel laminated beds in the upper part.

The section at location 87RH222-225 is not very different from the section above (Fig. 3.40). This section starts with a conglomeratic coarse-grained sandstone overlying a basal scour surface (Fig. 2.25). The sandstone becomes finer upward and is abruptly overlain by 3 m of coarse-grained sandstone which probably represents rapid deposition under high flow conditions. A further 5 m thick channel deposit overlies a basal scour surface. It consists of 1.30 m of massive sandstone followed by well bedded (40-100 cm thick) coarse-grained sandstone. The top three fining upward sequences represent channel-fill deposits containing slump structures, planar and trough cross-beds, and parallel laminated beds. Grain size and structures in the sandstone indicate that the lower part of this section shows a proximal fluvial character whereas the upper part represents lower energy depositional conditions.

A section through the upper part of the Tutoop Sandstone at location 87RH216 (Fig. 3.40) shows an overall finer grained sequence with thinner fining upward units. This section starts with a 2 m thick channel deposit overlying a basal scoured surface. The sandstone at the base is pebbly
and coarse-grained but it passes upward into well bedded (10-20 cm thick) fine- to medium-grained sandstone. The succeeding fining upward sequences are predominantly medium- to fine-grained and consist of planar and trough cross-beds. Individual sets are separated by scour surfaces. This section of multistory fining upward sandstone bodies probably represents fluvial channel deposits but they have a lower energy, more distal character compared with the sequences forming the bulk of the Tutoop Sandstone.

3.1.3c KETUNGAU FORMATION

Sedimentary structures are rare in the Ketungau Formation. They include planar cross-beds, and parallel and carbonaceous laminae. Coal seams are also present, especially in the middle of the formation.

Planar cross-beds are found within the sandstone beds (Fig. 3.42) where the average set thickness is about 0.22 m and the maximum thickness is 0.29 m. Flow depths for the production of these cross-beds could have ranged between 0.44 m and 2.77 m (Table 3.1). Using Froude numbers of 0.3 and 0.8, the flow velocities would have ranged from 0.45 m/s to 4.17 m/s with most estimates being less than 1 m (Table 3.1). Wavelengths of the dunes responsible for the deposition of the cross-bedded sandstone are estimated to range from 0.32 m to 24.93 m (Table 3.1).

Parallel and carbonaceous laminae are found within siltstone beds, and some of them exhibit slump folding (Fig. 3.41). The siltstone sequence probably represents
vertical accretion deposits that accumulated in an overbank floodplain or floodplain lake environment.

The distribution of sedimentary structures in the type section of the Ketungau Formation is shown in Fig. 3.42. However this unit shows considerable lateral variation in lithology and local abundance of sedimentary structures. A general feature in this formation is the decrease in number and thickness of sandstone units from the base to top of the formation.

The section at location 87RH205-207 is representative of the lower Ketungau Formation. The lower part of the section (about 11.5 m thick) is dominantly composed of mudstone with some intercalations of fine-grained sandstone. Bedding thickness ranges from 10 cm up to 50 cm. Some of the mudstone beds contain parallel lamination (e.g. Fig. 2.26), whereas planar and trough cross-stratification, and some parallel lamination is present in the sandstone beds. A few sandstone beds exhibit slump folding (e.g. Fig. 3.41). The dominance of mudstone in this part of the sequence indicates a low energy depositional environment while the presence of sandstone intercalations indicates that periodic higher energy conditions influenced this area. The most suitable depositional environment for this part of the sequence is a floodplain with the sandstone beds being derived from channel margin or crevasse splays. The upper part of the section consists of a 5 m thick well bedded (20-50 cm thick) medium-grained sandstone unit that erosionally overlies the mudstone. No sedimentary structures were recorded from this
upper sandstone unit, but in another area a similar sandstone unit contained some planar cross-beds. The depositional environment for this sandstone is uncertain, but it may represent shallow channelized flow or possibly a tidal flat or beach deposit.

The section at location 87RH233 from the central part of the Ketungau Formation is generally similar to the lower part of the previous section. It consists dominantly of mudstone with some intercalations of fine-grained sandstone. The sandstone is commonly parallel laminated. The presence of a thin coal seam indicates a fresh water influence and probably represents a backswamp environment on an extensive floodplain.

3.2 LITHOFACIES

The Melawi and Ketungau Basin sequences are composed of conglomerate, sandstone, siltstone and mudstone (shale). The bulk lithological proportions in each formation were determined by totalling the thickness of the appropriate grade of sediment in representative sections. The areas with no outcrop were counted as mudstone. The results are shown in Figures 3.43 and 3.44.

3.2.1 Melawi Basin Sequence

As can be seen in Figure 2.9, the Ingar Formation forms the lowest part of the Melawi Basin sequence in the studied area and consists of calcareous mudstone with some interbedded
siltstone and fine-grained sandstone. Limestone concretions are also found in the upper part of the Ingar Formation. The sandstone/mudstone ratio for this part of the Melawi sequence is about 1:4 (Fig. 3.43) and the lithological proportions are medium- to coarse-grained sandstone 5%, very fine- to fine-grained sandstone 10% and mudstone to siltstone 85%. The vertical and lateral distribution of lithofacies in the Ingar Formation are not clear, because of poor outcrop.

The middle part of the Melawi Basin sequence (Suwang Group) is dominantly composed of sandstone in the lower part (Dangkan Sandstone) and shale in the upper part (Silat Shale). Distribution of the Suwang Group in the studied area is shown in Figure 3.45.

The vertical distribution of lithofacies in the Dangkan Sandstone is shown in a stratigraphic column (Appendix 2), whereas the lateral distribution is shown in Figure 3.46. The Dangkan Sandstone consists of conglomerate (7%), very coarse-grained to pebbly sandstone (53%) and medium- to coarse-grained sandstone (40%; Fig. 3.43). As can be seen in Figure 3.46, the Dangkan Sandstone generally can be divided into two lithological subunits as follows: a well bedded fine- to medium-grained sandstone (subunit 1) and a massive to thick bedded coarse-grained to pebbly sandstone (subunit 2). In the western part of the basin (Dangkan River) the Dangkan Sandstone is dominated by massive to thick bedded coarse-grained to pebbly sandstone with the subunit 1 to subunit 2 ratio being 1:17. In an eastward direction the ratio increases to 1:3 in the middle part of the basin
(Nangangeri village) and 3:2 in the eastern part (Suwang River). The well bedded fine- to medium-grained sandstone subunit generally occupies the upper part of the formation.

The Silat Shale is dominantly formed of mudstone with some intercalations of siltstone and sandstone. The vertical distribution of lithofacies within the Silat Shale can be seen in Appendix 3, whereas the lateral distribution is shown in Figure 3.47. Generally the Silat Shale can be divided into two lithological subunits. Firstly, a shale subunit in the lower part of the formation, which dominantly consists of mudstone (shale) with minor intercalations of fine- to medium-grained sandstone. This subunit has a sand/shale ratio of about 1:17. Secondly, an interbedded sandstone-shale subunit in the upper part of the formation where the sand/shale ratio is about 5:8. The Silat Shale succession is well developed in the eastern part of the area, whereas in the western part of the basin the upper part of this formation was eroded before the deposition of the Melawi Group.

The upper part of the Melawi Basin sequence is composed of intercalated sandstone, siltstone and mudstone in the Melawi Group and dominantly sandstone in the overlying Kapuas Group.

Distribution of the Melawi Group in the studied area can be seen in Figure 3.48. The lateral and vertical lithological distribution of the Melawi Group is shown in Figure 3.49. In the northern part of the basin the Melawi Group consists dominantly of sandstone (Payak Formation) in the lower part and mudstone (Tebidah Formation), which
contains sporadic fossils, in the upper part. The sand/shale ratio for the lower formation is about 5:2, whereas the upper formation has a ratio of about 7:10. In the southern part of the basin the Melawi Group consist of interbedded mudstone, siltstone and sandstone with some coal seams in places. The thickness is up to 25 cm. Gastropods and bivalves are found in places within these sedimentary rocks and are a typical feature of the Tebidah Formation.

The vertical distribution of lithofacies within the Payak Formation is shown in a stratigraphic column (Appendix 4), whereas the lateral distribution is shown in Figures 3.49, 3.50, 3.52, 3.53 and 3.54. As can be seen in Figure 3.43 the Payak Formation consists of very coarse-grained to pebbly sandstone (5%), medium- to coarse-grained sandstone (15%), very fine- to fine-grained sandstone (45%) and mudstone to siltstone (35%). This formation thins out towards the south and it is correlated with Sepauk Sandstone and Shale 3 of the Tebidah formation on the southern limb of the Melawi Group (Figs 3.52 to 3.54).

As measured along the Mentatai River (Appendix 5), the Melawi Group consists of conglomerate (2%), very coarse grained to pebbly sandstone (6%), medium- to coarse-grained sandstone (27%), very fine- to fine-grained sandstone (10%) and mudstone to siltstone (55%). In the southern limb of the basin the Melawi Group can be divided into five lithological subunits as follows from top to bottom: Shale 1, Sandstone 1, and Shale 2 (Tebidah Formation), Sandstone 2 and Shale 3 (Figs 3.49 and 3.51). The thinnest sequence of Shale 1 is
found in the western part of the basin (Sepauk River), the unit progressively thickens northwards to the Kayan River and shows slight increase eastwards to the Mentatai River. Some hydrocarbon exploration companies in West Kalimantan used the term Lebang Shale for this subunit (B. Porthault, pers. comm., 1987), whereas the type locality of the Lebang (Van Emmichoven, 1939; Marks, 1957) is located in Lebang River and forms part of the Ingar Formation (Table 2.1).

Sandstone 1 is well developed in the western part of the area (Sepauk River), but it becomes slightly thinner northwards to the Kayan River. Eastwards Sandstone 1 shows a slight decrease in thickness towards the Mentatai River, but it then rapidly increases in thickness and is well developed in the Serawai River area.

Shale 2 is well developed in the southern part of the basin, especially at the Mentatai River. To the west it becomes slightly thinner towards the Sepauk River, whereas to the east it thins rapidly towards the Serawai River.

The Sepauk Sandstone is well developed in the western part of the area (Sepauk River) and becomes slightly thinner eastwards to the Pinoh River, and then rapidly decreases toward the Serawai River. To the north the Sepauk Sandstone thins out and may interfinger with the Payak Formation.

Shale 3 appears in the southeast corner of the area (Serawai River), its thickness decreases slowly westward and it wedges out in the Pinoh River. To the north Shale 3 is correlated with the Payak Formation. The Shale 3 is well developed in the eastern part of the basin where it is mapped.
as part of the Payak Formation (Fig. 2.1). Hydrocarbon exploration companies used the term Jengonoi Shale for this subunit (B. Porthault, pers. comm., 1987).

The Melawi Group is covered by the Kapuas Group; the Alat Sandstone in the eastern part of the basin and Sekayam Sandstone in the western part. The Alat Sandstone is well developed near Mt Alat. The vertical distribution of lithofacies within the Alat Sandstone is shown in Appendix 6. As shown in Figure 3.43, the Alat Sandstone consists of conglomerate (15%), very coarse-grained to pebbly sandstone (55%), medium- to coarse-grained sandstone (13%), and very fine- to fine-grained sandstone (17%). The Alat Sandstone can be divided into three lithological subunits. The lower part of the Alat Sandstone is composed of massive to thinly bedded very coarse-grained to pebbly sandstone interbedded with lenses of conglomerate. The middle part consists of massive to thinly bedded medium- to very coarse-grained sandstone with a few beds of conglomerate, whereas the upper part is formed of well bedded very fine- to fine-grained sandstone.

The Sekayam Sandstone is well developed in the western part of the area (Sanggau Quadrangle). As can be seen in Figure 3.43, it consists of very coarse-grained to pebbly sandstone (10%), medium- to coarse-grained sandstone (40%), very fine- to fine-grained sandstones (13%) and mudstone to siltstone (37%).
3.2.2. Ketungau Basin Sequence

The stratigraphic succession in the Ketungau Basin is shown in Figure 2.10, and its distribution in the studied area is shown in Figure 3.56.

The lower part of the sequence consists of interbedded quartzose, feldspathic and micaceous sandstone, carbonaceous siltstone and mudstone (Kantu Formation). The vertical distribution of lithofacies within the Kantu Formation is shown in Appendix 7, whereas the lateral distribution is shown in Figure 3.57. Generally the Kantu Formation can be divided into three subunits (Fig. 3.57). The lower part (675 m) consists dominantly of medium- to coarse-grained quartzlithic sandstone with a few conglomeratic beds. The continuation of this subunit to the northwest in Sarawak is called the Basal Sandstone Member of the Silantek Formation (Tan, 1979). The middle part of the Kantu Formation (2391 m) is occupied by intercalations of fine-grained sandstone, siltstone and mudstone, some of which are carbonaceous. Lenses of well bedded very fine- to fine-grained sandstone and well bedded medium- to coarse-grained sandstone are found in the Kantu and Kantuaping River areas (Fig. 3.56). The upper part of this formation (832 m) consists of bedded fine- to medium-grained sandstone with interbeds of red mudstone.

The lithological proportions in the Kantu Formation are shown in Figure 3.57. Conglomerate (2%) and very coarse-grained to pebbly sandstone (5%) occur in the lower part of the formation. Medium- to coarse-grained sandstone (16%) is dominant in the upper part of the formation but some
is also present in the middle and the lower parts of the formation. Very fine- to fine-grained sandstone (15%) occurs in the upper and middle parts of the formation. Mudstone and siltstone form the largest portion of the formation (62%) and are the dominant lithologies in the middle of the formation.

The Tutoop Sandstone in the middle of the Ketungau Basin sequence consists of massive to thickly bedded quartz-lithic arenite with some beds of conglomerate. The vertical distribution of the lithofacies within this unit is shown in Appendix 8, whereas the lateral distribution is shown in Figure 3.58.

The Tutoop Sandstone can be divided into two subunits (Fig. 3.58). The lower part is dominated by very coarse-grained to pebbly sandstone with some conglomerate beds. In the Segubung River area a lens of well bedded fine-grained sandstone with interbedded medium-grained sandstone is found in the middle of this subunit, whereas in the Ara River area a similar lens is found in the upper part of the subunit. The thickness of this subunit, which has been measured in the Segubung River, is 600 m but it decreases towards the west in the Ara River area. The upper part of the Tutoop Sandstone consists dominantly of medium- to coarse-grained sandstone. From the Segubung River the grain size becomes finer westward with the appearance of some very fine- to fine-grained sandstone and mudstone beds in the Ara River area. The thickness of the upper part the Tutoop Sandstone at Segubung River is about 900 m.
Lithological proportions in the Tutoop Sandstone at the Segubung River consist of conglomerate (8%), which occurs in the lower part of the formation; very coarse-grained to pebbly sandstone (18%) dominantly occurs in the lower part of the formation with some in the upper part; medium- to coarse-grained sandstone (65%) is dominant in the middle and upper parts of the formation; and very fine- to fine-grained sandstone (9%) is found in both the lower and upper parts of the formation.

The upper part of the Ketungau Basin sequence is formed of sandstone, siltstone and mudstone (Ketungau Formation). The vertical distribution of lithofacies within this formation can be seen in Appendix 9. Based on the four measured sections (Fig. 3.59) the Ketungau Formation shows considerable lateral variation. At the Sekapat River in the eastern part of the study area the Ketungau Formation is dominated by mudstone, with the sandstone to mudstone ratio being 2:3. The sandstone to mudstone ratio decreases upwards through the formation. Towards the west in the Ara River area the lithological proportions change dramatically with the appearance of sandstone and conglomerate beds. The sandstone+conglomerate to mudstone ratio increases to 6:5. West of the Ara River the sandstone to mudstone ratio decreases gradually to 9:10 at the Sekalau River and 3:5 in the upstream part of the Ketungau River.

Lithological proportions in the Ketungau Formation in the Sekapat River area (Fig. 3.44) consist of medium- to coarse-grained sandstone (7%), very fine- to fine-grained
sandstone (10%) and mudstone plus siltstone which forms the biggest portion (83%).
CHAPTER 4
PETROGRAPHY

4.1. PETROGRAPHIC METHODS

The ultimate goal of modal analyses of sandstone from the Melawi and Ketungau Basins is the reconstruction of original sandstone composition so that inferences concerning source terrains may be made. Each sample was point counted in order to increase the accuracy and internal consistency of the work. The maximum possible grid spacing was chosen so that 500 points with a spacing of 0.5 mm could be counted on each blue dye impregnated thin section.

The sandstone classification used in this study is based on the ternary system of Folk (1980) which is independent of textural features. The system is based on the proportion of the detrital components quartz (Q; which includes all type quartz plus metaquartzite), feldspar (F; which includes all single feldspar together with granite and gneiss fragments) and rock fragment (R; which includes all supracrustal rock fragments such as chert, limestone, sandstone, shale, slate, schist and volcanic detritus).

This study also used the standard petrographic techniques suggested by Ingersoll (1978, 1983) and Korsch (1984). The methods of data manipulation and analysis followed those outlined by Dickinson and Suczek (1979), Ingersoll and Suczek (1979), Dickinson (1982, 1985) and Dickinson et al. (1983a and 1983b).

For distinguishing lithic types, matrix types and other
components of each sandstone the criteria of Dickinson (1970) and Graham et al. (1976) were used. Monocrystalline grains larger than 0.0625 mm that occur within lithic fragments are counted with Q (quartz), F (feldspar) or M (monocrystalline phyllosilicate grains; after Dickinson, 1970).

For this study all point counts were recalculated as volumetric proportions for the following categories using the grain parameters defined by Ingersoll and Suczek (1979), and Dickinson and Suczek (1979): QFL, QmFLt, QpLvmLsm and LmLvLs. Where Q is total quartzose grains, Qm is monocrystalline quartz, Qp is polycrystalline quartz, F is total feldspar, P is plagioclase, K is K-feldspar, L is total unstable lithic fragments, Lm is metamorphic aphanitic grains, Lv is volcanic-hypabyssal aphanitic grains, Ls is sedimentary aphanitic grains, Lvm is volcanic-hypabyssal and metavolcanic, Lsm is sedimentary and metasedimentary aphanitic grains. In combined form: Q = Qm + Qp; F = P + K; L = Lm + Lv + Ls; and Lt = L + Qp. Extraneous constituents, such as heavy minerals and calcareous grains, are disregarded in this scheme (Dickinson and Suczek, 1979).

4.2 PETROGRAPHIC RESULTS

Point-count results are tabulated in Appendix 11, whereas the recalculated result are given in Table 4.1. Eleven parameters were used in this study. They include Q (total quartzose), Qm (monocrystalline quartz), Qp (polycrystalline quartz), F (feldspar), L (total unstable
aphanitic lithic grains), Lt (total aphanitic lithic grains), Lvm (volcanic-hypabyssal and metavolcanic aphanitic lithic grains), Lsm (sedimentary and metasedimentary aphanitic lithic grains), Lm (metamorphic aphanitic lithic grains), Lv (volcanic hypabyssal lithic grains), and Ls (sedimentary aphanitic lithic grains). The other calculated parameter is R (all supracrustal rock fragments including chert and limestone), which has been used for classification of the sandstone after Folk (1980). Means and standard deviations of each parameter for each formation are presented in Tables 4.2, 4.3 and 4.4.

Most feldspar in the studied area has been altered, therefore, plagioclase and potassium feldspar are very difficult to distinguish from each other and have been combined as a single category.

4.2.1 Petrological Components in the Melawi and Ketungau Basins

One hundred and ninety five thin-sections of sandstone from the Melawi and Ketungau Basins have been studied and tabulated in Appendix 11. The important mineralogical features are discussed below.

4.2.1a QUARTZ

Quartz is the most abundant mineral in sandstone from all formations in both the Melawi and Ketungau Basins, especially in the Dangkan, Alat and Tutoop Sandstones (Appendix 11).
The following features of quartz were recorded in this study: (1) single grains with straight extinction or extinction within less than two degrees rotation (Q. normal; Appendix 11); (2) single grains with undulose but continuous extinction with an extinction angle greater than two degrees; (3) single grains containing some inclusions (Fig. 4.1); (4) single grains (euhedral shape) with straight extinction and embayments (Fig. 4.2); (5) single grains with very abundant vacuoles (Fig. 4.3); and (6) single grains with warped, subparallel lines of small bubbles which have been termed Boehm lamellae. Other features in quartz grains consist of two or more (composite) interlocking quartz crystals with distinct straight (Fig. 4.4a) or undulose extinction and crenulated crystal boundaries (Fig. 4.4b).

Single grains of quartz with straight extinction are the most abundant constituents in all the sandstone samples studied (Appendix 11), specially within the Dangkan, Alat and Tutoop Sandstones where the average content varies between 48% and 61%. Quartz grains with straight extinction are generally subrounded to rounded with moderate to high sphericity. Quartz which contains vacuoles is usually the second most abundant quartz type but the average content is rare. These grains were probably derived from hydrothermal-vein sources. The other single quartz grains are very rare. Quartz grains with undulose extinction and/or inclusions are present within the Dangkan Sandstone. These grains indicate derivation from a metamorphic source area.

Embayed quartz with straight extinction indicates an
extrusive igneous source. Detrital quartz grains containing Boehm lamellae, are the product of intense strain deformation of quartz grains. Some single quartz grains have earlier generation overgrowths (Fig. 4.5).

Semicomposite and composite quartz grains form a minor part of the quartz assemblage; the maximum average content is found in the Dangkan Sandstone (22.0%), whereas the minimum average content is in the sandstone from the Ingar Formation (4.40%; Appendix 11). The grain shape varies from subangular to subrounded with moderate to high sphericity. Some of the grains contain a few acicular inclusions (Fig. 4.6). These grains are probably first cycle derivatives from a gneissic or granitic terrain.

4.2.1b FELDSPAR

Feldspar appears to be a major constituent throughout the sandstone from the Ingar (29.3%) and Tebidah (19.7%) Formations (Appendix 11). Most feldspar in sandstone units from the studied area has already been altered to clay minerals. Therefore, it is very difficult to distinguish between K-feldspar and plagioclase (recorded as undifferentiated feldspar; Fig. 4.7), but generally the K-feldspar content appears to be higher than plagioclase. The grains have a subangular to subrounded shape with moderate to low sphericity.

In the Dangkan Sandstone plagioclase grains are generally broken (Fig. 4.8) which indicates that the sandstone is deformed. In some places K-feldspar (Figs 4.9
and 4.10) and plagioclase have been partly or completely replaced by sericite and kaolinite, and some grains have overgrowths.

4.2.1c ROCK FRAGMENTS

Rock fragments are generally present in all samples of sandstone from the studied area. The most abundant rock fragment constituent is volcanic detritus, especially in the Sekayam Sandstone (30.3%) and in the sandstone from the Ingar Formation (23.1%). The rest of the formations contain less than 16% of rock fragments (Appendix 11).

Volcanic rock fragments are present as subangular to subrounded grains with moderate to high sphericity. Glass shard structures still can be recognized (Fig. 4.11), as well as microlites of feldspar (Fig. 4.12). Some volcanic rock fragments have been partly dissolved (Fig. 4.13) and replaced by sericite and kaolinite. In the Dangkan Sandstone some volcanic fragments are replaced by laumontite. Felsic igneous fragments are found within the Dangkan Sandstone and are marked by a micrographic texture within the grain (Fig. 4.14). Ultramafic rock fragments are also found within the Dangkan Sandstone (Fig. 4.15). The presence of felsic and ultramafic rock fragments indicate that both these constituents were derived from the Boyan Melange.

Metamorphic rock fragments form less than 2% of the framework grains and they are absent from many samples. These grains are generally subrounded to subangular with moderate to low sphericity. Most of the metamorphic
fragments consist of metaquartzite (Fig. 4.16) and they are most common in the Dangkan Sandstone. These grains indicate derivation from a metamorphic source.

Sedimentary rock fragments, which form a minor component in the studied area, consist of sandstone (Fig. 4.17), siltstone (Fig. 4.18) and limestone. The average content of these rock fragments in each formation is less than 1%. Sandstone fragments are more common than siltstone and limestone grains. They have a subangular to subrounded shape with moderate to low sphericity.

Chert rock fragments are also found within the sandstone from the studied area. The average content of chert grains is less than 9%, and they are subangular to subrounded with moderate to high sphericity.

4.2.1d MICAS

Detrital micas such as biotite and muscovite are relatively common in the sandstone in the studied area. Muscovite is more common than biotite, probably because it is more resistant to weathering. The average content of muscovite is less than 2% and for biotite it is less than 1%. The micas appear as subangular to subrounded flakes with moderate to low sphericity. In the Suwang Group both muscovite (Fig. 4.19) and biotite (Fig. 4.20) have been deformed.

Two types of muscovite are present in the sandstone in the studied area. Primary muscovite occurs as individual flakes, some of which have been deformed (Fig. 4.19), whereas
secondary muscovite occurs along cracks and cleavage planes in broken plagioclase grains (Fig. 4.9).

4.2.1e HEAVY MINERALS

Heavy minerals present within the sandstone in the studied area include epidote, tourmaline, opaque minerals, zircon, sphene, pyroxene and garnet. The average content of heavy minerals is less than 1% (Appendix 11).

Epidote, tourmaline, opaque minerals, zircon and sphene are the most common heavy minerals which are found in most formations. The main opaque mineral recognized is pyrite which occurs in two forms. Primary pyrite consists of subhedral or anhedral crystals whereas secondary authigenic pyrite appears as euhedral crystals (Fig. 4.21). Garnet is only found in a few samples and pyroxene is limited to samples from the Dangkan and Sekayam Sandstones.

4.2.1f OTHER ACCESSORY COMPONENTS

Other accessory components which are found within the sandstone in the studied area include glass, organic matter, siderite and glauconite. Generally they occur with an average content of less than 1% (Appendix 11). Glass is present as fine grains, with a subrounded to subangular grain shape and moderate to high sphericity. Organic matter was only found within sandstone beds from the Ingar Formation, Silat Shale and Tebidah Formation. Organic matter within the sandstone occurs either as grains or as carbonaceous laminae (Fig. 4.22). Glauconite was only recorded in sandstone from the Payak and Tebidah Formations.
4.2.1f MATRIX

The term detrital matrix is used to include all material finer than 35 microns. It includes fine detrital quartz grains, accessory and opaque minerals, and detrital and authigenic clay minerals such as kaolinite, sericite and chlorite. Generally the average content of matrix is less than 10% (Appendix 11), but it is present in most of the sandstone in the studied area. Clay minerals increase in abundance with increasing feldspar and volcanic rock fragment contents. Secondary clays may line intergranular pores or they may fill up or replace dissolved grains (Fig. 4.23). The average clay mineral content is very low within the sandstone which is dominated by quartz grains, such as in the Dangkan and Alat Sandstones in the Melawi Basin, and the Tutoop Sandstone in the Ketungau Basin.

4.2.1g CEMENT

Cementation is the principal process leading to porosity reduction in sandstone. The main cement components include carbonate and laumontite cement, quartz overgrowths and iron oxide cements.

The occurrence of carbonate cement, dominantly calcite, may be either primary or secondary. The primary cements consist of a mosaic of very fine-grained calcite. Such features are found within some sandstone beds in the Ingar, Payak and Tebidah Formations. Secondary carbonate cements include replacement of detrital grains by calcite and growth of authigenic calcite. These features are present in some of
the sandstone in the Ingar, Payak and Tebidah Formations, and also in the Dangkan Sandstone and Silat Shale. The highest average content of carbonate cement is found in the Ingar Formation (10%), while in the other formations it is usually less than 2% (Appendix 11).

Laumontite cement (Fig. 4.24) was only found within some sandstone samples in the Ingar Formation, Dangkan Sandstone and Silat Shale. The average content is less than 1%. The presence of laumontite indicates that the sandstone in these formations has undergone moderate burial diagenesis.

Siliceous cement and quartz overgrowths (Fig. 4.6) are present in the quartz-rich sandstone units such as the Dangkan, Alat and Tutoop Sandstones. The average content of quartz cement is less than 5%.

Iron oxides are also found within the quartz-rich sandstone units such as the Dangkan, Alat and Tutoop Sandstones, but the average content is less than 2%.

### 4.2.2 Petrological characteristics of each formation

The significance of the following petrological characteristics are assessed in Chapter 4.4.

#### 4.2.2a INGAR FORMATION

Ten thin sections have been studied from sandstone beds within the Ingar Formation (Appendix 11). Microscopically the sandstone is composed of more than 80% framework grains, with a maximum grain size of 0.5 mm and an average grain size of 0.1 mm. The sphericity of the grains is generally low, they have a subangular shape and are well sorted. The
maturity of these sandstone units ranges from submature to mature (Appendix 11). A typical thin section view of the sandstone can be seen in Figure 4.25.

The framework grains consist of 9-21% quartz, 1-16% composite quartz, 15-39% feldspar and 12-40% lithic fragments. According to the QFR sandstone classification (Folk, 1980; Fig. 4.26a,b) sandstone beds within the Ingar Formation range from lithic arkose to feldspathic litharenite. The means and standard deviations of each parameter are shown in Table 4.4. The unstable lithic fragments are dominated by volcanic rock (Fig. 4.30a,b). Other lithic fragments include limestone, which is found in some of the samples. Accessory minerals are mostly muscovite (up to 2.8%), with some chlorite (up to 1.2%). Heavy minerals are very rare, with no opaque minerals and tourmaline only being found in a few samples. The matrix generally consists of clay minerals which are dominated by kaolinite with some sericite and chlorite. Spar calcite occurs in some samples (Appendix 11).

The QFL, QmFLt and QpLvmLsm diagrams for these samples (Figs 4.27a, 4.28a and 4.29a), provide relevant means and standard deviations for each parameter (Figs 4.27b, 4.28b and 4.29b) as shown in Tables 4.2 and 4.3. Based on the QFL and QmFL diagrams (Figs 4.27 and 4.28), which can be compared with the QtFL diagram of Dickinson et al. (1983a) and Dickinson and Suczek (1979; Figs 4.78 and 4.79) the sandstone from the Ingar Formation was probably derived from a magmatic arc, ranging from a dissected arc to a transitional arc
(Dickinson et al., 1983a). According to the QpLvLsm diagram (Fig. 4.29; cf. the QpLvLs diagram of Dickinson and Suczek, 1979, Fig. 4.79) the Ingar Formation shows a very close relationship with an arc orogen source. However, if the diagram is compared with the triangular plot of Ingersoll and Suczek (1979; Fig. 4.80) the source may fall into the magmatic arc or forearc areas.

The mean and standard deviation of the Lv/L ratio for the sandstone within the Ingar Formation is $0.97 \pm 0.05$, whereas the mean and standard deviation of Qp/Q is $0.20 \pm 0.10$.

4.2.2b SUWANG GROUP

Twenty seven thin sections were made from the Dangkan Sandstone and 16 thin sections from sandstone within the Silat Shale. Petrographically the sandstone from these two formations shows clear differences (Appendix 11). Typical views of the Dangkan Sandstone and the sandstone from the Silat Shale are shown in Figures 4.31 and 4.32.

Most of the Dangkan Sandstone is white, whereas the sandstone from the Silat Shale is grey. The Dangkan Sandstone is composed of 86% to 100% framework grains, while the sandstone from the Silat Shale consists of 82% to 98% framework grains. The maximum and the average grain size in the Dangkan Sandstone are 2.5 mm and 0.92 mm, compared with 2.0 mm and 0.36 mm in the sandstone from Silat Shale. Grain sphericity from both formations are moderate, the shape ranges from subangular to subrounded and the sandstone is generally moderately sorted (Appendix 11).
Framework grains in the Dangkan Sandstone consist of 27.5-80% quartz, 11.7-42% composite quartz, 0-14.2% feldspar and 0-27% lithic fragments. The sandstone from the Silat Shale, however, consists of 22-69.5% quartz, 0.5-22.7% composite quartz, 1.7-44.2% feldspar and 0.5-23.4% lithic fragments. As an accessory mineral muscovite (0-1.7%) is present in both formations, whereas biotite is only present in the sandstone from the Silat Shale. The both formations also contain epidote (0-0.7%), tourmaline (0-0.7%) and opaque minerals (0-0.5%). The sandstone matrix from both formations is composed of clay minerals, dominated by kaolinite with some sericite and chlorite. Spar calcite also occurs as a cement within some samples. Organic material is only found in the sandstones from the Silat Shale (Appendix 11).

According to QFR diagram (Fig. 4.33a) the Dangkan Sandstone consists predominantly of sublitharenite, with the mean and standard deviation of each parameter being 88±10% quartz, 2±5% feldspar and 10±9% rock fragments (Table 4.4). Sandstone from the Silat Shale, however, falls into the lithic arkose to feldspathic litharenite fields, with the mean and standard deviation of each parameter being 54±16% quartz, 23±15% feldspar and 23±13% rock fragments (Table 4.4). As shown in Figure 4.37a, unstable rocks fragments within the Dangkan Sandstone include 18±24% metamorphic, 79±22% volcanic and 3±10% sedimentary lithic fragments (Table 4.3). Rock fragments within the sandstone from the Silat Shale consist of 2±3% metamorphic, 97±3% volcanic and 1±1% sedimentary lithic detritus (Table 4.3) with limestone fragments present in a few samples (Appendix 11).
The QFL, QmFLt and QpLvmLsm are shown in Figures 4.34, 4.35 and 4.36, whereas the means and standard deviations of each parameter can be seen in Tables 4.2 and 4.3. The QFL diagram for the Suwang Group sandstone (Fig. 4.34), when compared with diagrams of Dickinson and Suczek (1979) and Dickinson et al. (1983a; Figs 4.78 and 4.79), suggests a recycled orogenic provenance, whereas the QmFLt diagram (Fig. 4.35) indicates that the Dangkan Sandstone might have come from quartzose recycled area and the sandstone from the Silat Shale represents a mixed provenance area. The QpLvmLsm diagram (Fig. 4.36; cf Fig. 4.79) indicates that the Dangkan Sandstone falls between collision suture — fold-thrust belt and subduction complex sources, whereas the sandstone from the Silat Shale shows a close relationship with subduction complex sources. Comparison of Figure 4.36 and 4.80 (Ingersoll and Suczek, 1979) suggests that sandstone from the Suwang Group is related to a rifted continental margin source (eugeoclines and abyssal plain). The LmLvLs diagram (Fig. 4.37; cf. Fig. 4.80) indicates that sandstone from the Suwang Group represents forearc deposition derived from a magmatic area.

The mean and standard deviation of the Lv/L ratio for the Dangkan Sandstone is 0.76±0.30, while that of the sandstone from the Silat Shale is 0.98±0.20. The mean and standard deviation of Qp/Q for the Dangkan Sandstone is 0.30±0.11, whereas that of the sandstone from the Silat Shale is 0.29±0.14.
4.2.2c MELAWI GROUP

Seventeen thin sections from sandstone in the Payak Formation and 21 thin sections from the Tebidah Formation were made, but unfortunately, due to limited time and problems of access when the field work was undertaken, only four thin sections could be made to represent the Sepauk Sandstone. Typical views of the sandstone from the Payak and Tebidah Formations can be seen in Figures 4.38 and 4.39.

Petrographically sandstone from the Payak and Tebidah Formations are similar, whereas the Sepauk Sandstone shows differences in some variables. Most sandstone from this group is grey and the percentage of framework grains ranges from 90-100% for the Payak Formation, 85-98% for the Tebidah Formation and 95-100% for the Sepauk Sandstone. The maximum and the average sandstone grain sizes are 2.00 mm and 0.26 mm for the Payak Formation, 0.80 mm and 0.21 mm for the Tebidah Formation, and 2.00 mm and 0.37 mm for the Sepauk Sandstone. All the sandstone is submature to mature and is characterized by moderate grain sphericity, subangular to subrounded grain shape, and well sorted texture (Appendix 11).

The framework grains of the sandstone from the Payak Formation include quartz (28-56%), composite quartz (1-28%), feldspar (6-51%) and lithic fragments (5-19%). The Tebidah Formation consists of quartz (30-72%), composite quartz (2-19%), feldspar (3-36%) and lithic fragments (2-26%). The Sepauk Sandstone includes framework quartz (34-79%), composite quartz (4-23%), feldspar (2-15%) and lithic fragments (6-27%). The most common accessory minerals
present within sandstone from these formations are muscovite (0.5-2.5%) and chlorite (up to 1%). Biotite is only present in a few samples from the Tebidah Formation. Heavy minerals include tourmaline and epidote (up to 0.5%) and opaque minerals. Most of the matrix within the sandstone from these formations dominantly consists of kaolinite with some sericite and chlorite. Sparry calcite is also found as a cement in a few samples. Organic matter within the sandstone from these formations is only found in a few places with a content up to 1% (Appendix 11).

Based on the QFR classification diagram (Fig. 4.40a,b) sandstone from the Payak Formation is dominantly lithic arkose with the mean and standard deviation of each parameter being 52±10% for quartz, 26±10% for feldspar and 22±6% for rock fragments (Table 4.4). Sandstone from the Tebidah Formation mainly falls into the feldspathic litharenite field with the mean and standard deviation for each parameter being 53±10% for quartz, 22±8% for feldspar and 25±8% for rock fragments (Table 4.4). The Sepauk Sandstone is also mainly feldspathic litharenite with the mean and standard deviation for each parameter being 73±16% for quartz, 8±7% for feldspar and 19±9% for rock fragments. Unstable rock fragments present in these formations are dominated by volcanic detritus (Fig. 4.44a,b) with the mean and standard deviation of each parameter of the sandstone being 1±2% metamorphic, 98±3% volcanic and 1±1% sedimentary fragments for the Payak Formation; 3±5% metamorphic, 96±5% volcanic and 1±1% sedimentary fragments for the Tebidah Formation; and 3±3%
metamorphic, 96±4% volcanic and 1±1% sedimentary fragments for the Sepauk Sandstone (Table 4.3).

The QFL, QmFLt and QpLvmLsm diagrams for sandstone from the Melawi Group are shown in Figures 4.41, 4.42, and 4.43, with the means and standard deviations of each parameter listed in Tables 4.2 and 4.3. The QFL diagram (Fig. 4.41; cf. Figs. 4.78 and 4.79) indicates that sandstone from the Melawi Group probably came from a recycled orogenic provenance area. However, the QmFLt diagram (Fig. 4.42; cf. Fig. 4.78) for the same sequence indicates a mixed provenance area. The QpLvmLsm (Fig. 4.43; cf. Fig. 4.79) indicates that the sandstone from the Melawi Group was derived from subduction complex sources, whereas if compared with the diagram of Ingersoll and Suczek (1979; Fig. 4.80) it shows that a proportion of the sandstone from the Melawi Group falls beyond the field of any defined depositional setting. The LmLvLs diagram (Fig. 4.44; cf. Fig. 4.80) indicates that some of the sandstone from the Melawi Group was probably deposited in a forearc area adjacent to a magmatic arc.

The means and standard deviations of the Lv/L and Qp/Q ratios for the sandstones from these formations show only slight differences. The mean and standard deviation of Lv/L and Qp/Q for sandstone from the Payak Formation are 0.98±0.02 and 0.26±0.09, whereas for the Tebidah Formation they are 0.96±0.05 and 0.24±0.09, and for the Sepauk Sandstone they are 0.98±0.03 and 0.22±0.12 (Table 4.4).
4.2.2d KAPUAS GROUP

Fifteen thin sections were made from the Sekayam Sandstone and 26 thin sections were made from the Alat Sandstone. Typical views of the Sekayam and Alat Sandstones can be seen in Figures 4.45 and 4.46.

Petrographically the Sekayam and Alat Sandstones are quite different. Most of the Sekayam Sandstone is grey, whereas the Alat Sandstone ranges from white to grey and brownish colour (Appendix 11).

The Sekayam Sandstone consists of 89-97% framework grains, whereas the Alat Sandstone contains 90-100% framework grains. The maximum and average grain size of the Sekayam Sandstone are 1.3 mm and 0.49 mm, whereas from the Alat Sandstone they are 4.0 mm and 0.58 mm respectively. Generally, grain sphericity from the Sekayam Sandstone ranges from low to moderate, with the grain shape mostly subangular. The Alat sandstone, however, has moderate to high grain sphericity with the grain shape ranging from subangular to subrounded (Appendix 11).

Framework grains in the Sekayam Sandstone include 12-39% quartz, 4-29% composite quartz, 6-24% feldspar and 20-45% lithic fragments. Accessory minerals, such as muscovite (up to 1.8%), biotite (up to 0.2%) and chlorite (up to 1.2%) are only present in some samples. Epidote (up to 0.6%) is a common heavy mineral within this sandstone, whereas tourmaline, zircon and sphene only occur as very rare minerals (up to 0.2%). Framework grains in the Alat Sandstone consist of 28-76% quartz, 8-40% composite quartz,
1-23% feldspar and 3-40% lithic fragments. As in the Sekayam Sandstone, the most common accessory mineral within the Alat Sandstone is muscovite (up to 1.4%), whereas biotite and chlorite are rare. Heavy minerals within the Alat Sandstone include epidote, tourmaline, zircon and sphene (Appendix 11).

Clay matrix is found in both the Sekayam and Alat Sandstones and it is dominated by kaolinite. Silica cement is found in both sandstones, whereas iron oxide is found only within the Alat Sandstone. Sparite and micrite are present in some parts of the Sekayam Sandstone only (Appendix 11).

Based on the QFR classification diagram (Fig. 4.47a,b), the Sekayam Sandstone falls between feldspathic litharenite and litharenite, with the means and standard deviations being 49±10% for quartz, 13±8% for feldspar and 38±7% for rock fragments. The Alat Sandstone, however, is mainly sublitharenite with the means and standard deviations being 76±12% for quartz, 8±7% for feldspar and 16±9% for rock fragments (Table 4.4). Lithic volcanic fragments are the dominant unstable lithic component within both the Sekayam and Alat Sandstones (Fig. 4.51a,b). The means and standard deviations of each parameter in the Sekayam Sandstone are 5±7% metamorphic, 93±9% volcanic and 2±3% sedimentary lithic fragments (Table 4.3). Comparable means and standard deviations for the Alat Sandstone are 12±9% metamorphic, 85±9% volcanic and 3±5% sedimentary lithic fragments (Table 4.3).

The QFL, QmFLt and QpLvmLsm diagrams for both the Sekayam and Alat Sandstones (Figs 4.48, 4.49 and 4.50) have a
mean and standard deviation for each parameter as mention in Tables 4.2 and 4.3.

The QFL and QmFLt diagrams for sandstone from the Kapuas Group (Figs. 4.48 and 49), which can be compared with the diagrams of Dickinson and Suczek (1979) and Dickinson et al. (1983a; Figs. 4.78 and 4.79), indicate that the sandstone came from a recycled orogenic provenance area; the Alat Sandstone lies between quartzose and transitional recycled provenance areas, whereas the Sekayam Sandstone falls into the transitional recycled provenance area. The QpLvmLsm diagram (Fig. 4.50; cf. Fig. 4.79) shows that the Sekayam Sandstone most probably came from an arc orogenic source, whereas the Alat Sandstone is more closely related to a subduction complex source. However, comparison of Figure 4.50 and Figure 4.80 suggests that the Sekayam Sandstone may have been deposited in a back arc basin derived from mixed magmatic arc and rifted continental margin sources. The Alat Sandstone does not conform to any defined depositional setting on Figure 4.80. The LmLvLs diagram for sandstone from the Kapuas Group (Fig. 4.51; cf. Fig. 4.84) suggests a forearc depositional setting for the Kapuas Group with detritus from a magmatic arc.

The mean and standard deviation of the Lv/L ratio for the Sekayam Sandstone is 0.93±0.08 compared with 0.85±0.09 for the Alat Sandstone. The mean and standard deviation of the Qp/Q ratio for the Sekayam Sandstone is 0.36±0.18, compared to 0.33±0.15 for the Alat Sandstone.
4.2.2e MERAKAI GROUP

Twenty two thin sections were made from sandstone in the Kantu Formation, 23 from the Tutoop Sandstone and 14 from the Ketungau Formation. Typical views of the sandstone from Merakai Group are shown in Figures 4.52, 4.53 and 4.54.

Petrographically the sandstone from each of these formations is clearly different. Most of the Tutoop Sandstone is white, whereas sandstone from the Kantu and Ketungau Formations is grey. Sandstone from the Kantu Formation contains 87-97% framework grains, whereas the Tutoop Sandstone consists of 93-100% framework grains and the sandstone from the Ketungau Formation has 83% to 98% framework grains. The maximum and the average grain size for each formation are 0.80 mm and 0.33 mm for the Kantu Formation, 1.2 mm and 0.40 mm for the Tutoop Sandstone, and 0.80 mm and 0.24 mm for the Ketungau Formation. Sandstone grains from the Kantu Formation have a moderate to high sphericity and are subangular to subrounded, whereas those from the Tutoop Sandstone and Ketungau Formation have moderate sphericity and subangular to subrounded shapes. Most sandstone from these formations is well sorted (Appendix 11).

Framework grains in sandstone from the Kantu Formation consist of 24-40% quartz, 8-25% composite quartz, 8-23% feldspar and 7-26% lithic fragments. The Tutoop Sandstone contains 50-70% quartz, 4-31% composite quartz, up to 7% feldspar and 7-14% lithic fragments. Sandstone from the Ketungau Formation is composed of 42-61% quartz, 9-30%
composite quartz, 1-9% feldspar and 4-20% lithic fragments.

As an accessory mineral muscovite is common in sandstone from the Kantu Formation (0.6-2.8%), whereas in the Tutoop Sandstone and Ketungau Formation it is rare (up to 0.5%). Biotite is very rare within these formations. Epidote, tourmaline and opaque minerals are common heavy minerals within these formations, whereas zircon, sphene and garnet are very rare. Organic matter is very rare within both the Kantu and Ketungau Formations, whereas in the Tutoop Sandstone it is absent (Appendix 11).

Clay matrix is present within all these formations. The content in sandstone from Kantu and Ketungau Formations is higher (up to 10%) than that of the Tutoop Sandstone (up to 5%). Iron oxide cement is common within the Tutoop Sandstone (up to 5%), but is absent from the Kantu and Ketungau Formations (Appendix 11).

According to the QFR diagram (Fig. 4.55a,b) sandstone from the Kantu Formation falls into the feldspathic litharenite field with the mean and standard deviation for each parameter being 59±7% quartz, 14±7% feldspar and 27±5% rock fragments. The Tutoop Sandstone is mainly sublitharenite with the mean and standard deviation for each parameter being 82±5% quartz, 3±2% feldspar and 15±4% rock fragments. Sandstone from the Ketungau Formation ranges between sublitharenite, feldspathic litharenite and litharenite, with the mean and standard deviation of each parameter being 74±8% quartz, 7±3% feldspar and 19±6% rock fragments (Table 4.4).
Lithic volcanic detritus is the dominant unstable rock fragment within all of these formations (Fig. 4.59a,b). The mean and standard deviation of each parameter in the sandstone are 7±5% metamorphic, 92±5% volcanic and 1±1% sedimentary lithic fragments for the Kantu Formation; 10±4% metamorphic, 85±5% volcanic and 5±3% sedimentary lithic fragments for the Tutoop Sandstone; and 8±4% metamorphic, 86±6% volcanic and 6±5% sedimentary lithic fragments for the Ketungau Formation (Table 4.3).

The QFL, QmFLt and QpLvmLsm triangular diagrams for the Merakai Group (Figs 4.56, 4.57 and 4.58) have means and standard deviations of each parameter mentioned in Tables 4.2 and 4.3. The QFL and QmFLt diagrams (Figs 4.56 and 4.57; cf. Figs 4.78 and 4.79) indicated a transitional recycled orogen provenance for the Kantu and Ketungau Formations, whereas the Tutoop Sandstone consist of detritus from a quartzose recycled orogen provenance. The QpLvmLsm diagram (Fig. 4.58; cf. Fig. 4.79) suggests a subduction complex source for sandstone from the Merakai Group. However, comparison of Figure 4.58 and Figure 4.80 indicates that a proportion of sandstone from the Merakai Group plots beyond the field of any defined depositional setting. According to the LmLvLs diagram (Fig. 4.59; cf. Fig. 4.80) the sandstone from the Merakai Group may represent forearc deposits derived from magmatic arc.

The mean and standard deviation of the lithic volcanic/ unstable lithic fragment ratio and polycrystalline quartz/ total quartz ratio for sandstone from the Kantu Formation are
0.93±0.05 and 0.40±0.11, whereas equivalent ratios for the Tutoop Sandstone are 0.84±0.04 and 0.23±0.07, and for sandstone from the Ketungau Formation are 0.86±0.05 and 0.30±0.06.

4.3. CLUSTER ANALYSIS

An analysis of the petrological data from sandstone units in the Melawi and Ketungau Basins was undertaken using correlation coefficient and cluster analysis. The program used was based on Davis (1973) and Jones and Facer (1982). Computations were performed using the program devised by Dr B.G. Jones (pers. comm., 1990) and the main data file is presented in Appendix 11.

Correlation coefficients were chosen as a technique for analyzing data as they were designed to analyze large data matrices with interdependence (Hair et al., 1979). The correlation coefficient is used as a measure of similarity between variables or samples. Q-mode analysis is used to calculate the correlation coefficients between samples, whereas R-mode analysis calculates the correlation coefficients between the variables. The similarity coefficients thus produced can be grouped using cluster analysis. The highest similarities are clustered or linked first. The two samples or variables with the highest similarity coefficient are linked together and their correlation with all other samples are averaged. As this process proceeds the relationship between all samples or
variables can be presented by a dendritic network or dendrogram.

In the Q-mode analysis a cosine-theta similarity coefficient was used which expresses the similarity as an angular relationship between each two samples in a multidimensional coordinate system (Harbaugh and Merriam, 1968). The angular separation defined may range from 0°, indicating complete similarity with a coefficient of 1.0, to 90°, corresponding to uncorrelated data with coefficient of 0.0, to 180°, corresponding to a completely negative relationship with coefficient of -1.0. In the R-mode analysis a Pearson Product Moment correlation coefficient was used for comparisons between numerical variables. This correlation coefficient may be visualized as the degree to which observations of two variables approach a straight line on an X-Y diagram, and it may assume a value ranging from -1.0 to +1.0. Zero indicates no linear correlation while positive values indicate positive correlation and negative values indicate negative correlation (Harbaugh and Merriam, 1968). The correlation coefficients were then tested for significance and a matrix of significant correlation coefficients were plotted at 95% and 99% confidence levels.

The data for each analysis consisted of up to one hundred samples each with fifty eight variables. As the data for petrographic textures are non-numeric, a coded numeric variable as shown in Appendix 11 was used. The numeric value for colour increased from 1 (white) to 5 (black). Packing increased from very poorly to very well packed and porosity
increased from very low to very high, both with numeric values ranging from 1 to 5. Sphericity of the grains increased from low to high and sorting increased from poorly sorted to well sorted sandstone, both having a numeric values ranging from 1 to 3. Roundness increases from very angular to well rounded grains having values ranging from 1 to 6. The numeric value for maturity increased from 1 for immature sandstone to 4 for supermature sandstone. Variables grouped during point counting, such as heavy minerals, were assigned a numeric value of 1 for very rare to 4 for abundant.

4.3.1 Q-mode Analysis

This analysis is especially useful for comparison between sedimentary facies and mineral assemblages (Harbaugh and Merriam, 1968). Ten Q-mode dendrograms were made in order to compare the similarity of the sandstone within and between each formation, each group and each basin.

4.3.1a SUWANG GROUP

Forty three sandstone samples from the Suwang Group, each with 58 variables (Appendix 11), were used to calculate correlation coefficients between these samples (Fig. 4.65). Based on this dendrogram sandstone from the Suwang Group can be divided into three major divisions.

The first group consists of 12 samples from the Dangkan Sandstone with correlation coefficient ranging from 0.1028 to 0.6495. This group represents a submature sandstone which is characterized by moderate packing, poor to moderate sorting,
and subangular grain shape with low to moderate sphericity. These sandstone samples contain less than 60% quartz.

The second group contains 15 samples mainly consisting of sandstone from the Silat Shale with correlation coefficients ranging from 0.1291 to 0.7353. This group is very poorly correlated with the previous group (correlation coefficient = 0.0139). The second group consists of submature to mature sandstone samples which are moderately to very well packed, poorly to well sorted, and have subangular to subrounded grain shapes with low to moderate sphericity. Mineralogically this group is characterized by less than 45% quartz and up to 44% feldspar.

These two groups have a slightly negative correlation (-0.0096) with the third group that consists of 16 samples mainly from the Dangkan Sandstone. This group has correlation coefficients ranging from 0.0725 to 0.7698 and it represents mature sandstone which is characterized by moderate to good sorting, subrounded to rounded grain shape with moderate sphericity, and the samples are all well to very well compacted. Mineralogically this group is characterized by high contents of quartz (up to 60%) and low contents of feldspar (up to 14%).

4.3.1b MELAWI GROUP

Forty two samples from the Melawi Group, each with 58 variables (Appendix 11), were used to determine the petrographic relationships between sandstone samples from the Payak and Tebidah Formations and the Sepauk Sandstone (Fig.
The Q-mode dendrogram (Fig. 4.66) for sandstone from the Melawi Group shows no divisions that represent any formation. Each division comprises a mixture of sandstone from the three formations. The samples are all positively correlated with the coefficients ranging from 0.1781 to 0.8489.

The presence of positive correlation between each sample, and the mixture of samples from all formations in each cluster group, indicates that based on the Q-mode dendrogram (Fig. 4.66) the sandstone samples from the Melawi Group can not be separated. Thus sandstone samples from the Payak and Tebidah Formations and the Sepauk Sandstone have a close relationship to each other. The similarity of the sandstone from these formation is also indicated by the average volcanic fragment content, and the \(Lv/L\) and \(Qp/Q\) ratios. The average volcanic fragment contents are 12.3\% for the Payak Formation, 15.2\% for the Tebidah Formation and 12.2\% for the Sepauk Sandstone (Appendix 11). The \(Lv/L\) and \(Qp/Q\) ratios are 0.98±0.02 and 0.26±0.09 for the Payak Formation, 0.96±0.05 and 0.24±0.09 for the Tebidah Formation, and 0.98±0.03 and 0.22±0.12 for the Sepauk Sandstone (Table 4.4).

4.3.1c KAPUAS GROUP

The Q-mode dendrogram (Fig. 4.67) relationship between sandstone samples from the Kapuas Group was determined from 41 samples each with 58 variables (Appendix 11).

Based on the Q-mode cluster analysis (Fig. 4.67), the
sandstone samples from the Kapuas Group can be divided into two major divisions. The first division consists of 28 samples dominated by the Alat Sandstone. The correlation coefficients between samples in this division range from 0.0317 to 0.7082. The Alat Sandstone forming this division is characterized by a high content of quartz (48.6% average, ranging from 28.4% to 76.0%) and composite quartz (21.4% average, ranging from 4.8% to 40.0%), and low contents of feldspar (6.6% average, ranging from 1.0% to 22.8%) and volcanic rock fragments (9.9% average, ranging from 2.6% to 39.0%; Appendix 11). Three samples from the Sekayam Sandstone are included in this division (83PR11A, 83BA24A and 83BH19A) because they contain 31.6-31.8% quartz, 14.6-18.9% composite quartz and 5.6-11.4% feldspar.

The correlation coefficient between the first and second division is 0.0056. The second division consists of 13 samples, dominated by the Sekayam Sandstone with, correlation coefficients between samples ranging from 0.0892 to 0.7587 (Fig. 4.67). This division represents the Sekayam Sandstone and is characterized by a lower content of quartz (28.9% average, ranging from 12.4% to 39.0%) and composite quartz (15.9% average, ranging from 3.8% to 48.0%), and a higher content of feldspar (11.2% average, ranging from 0.8% to 23.8%) and volcanic rock fragments (30.3% average, ranging from 18.8% to 44.2%). Only one sample from the Alat Sandstone is included this division (87RH24B) since it has a lower quartz content (30.00%) and a higher volcanic rock fragment content (18.00%) than the other samples from the
4.3.1d MERAKAI GROUP

The Q-mode dendrogram of sandstone from the Merakai Group (Fig. 4.68) was based on 59 samples, each with 58 variables (Appendix 11). Based on this dendrogram, sandstone samples from the Ketungau Basin can be divided into two major divisions.

The first division (Group I) consists of 22 sandstone samples from the Kantu Formation, which are characterized by a grey colour, are moderately to well compacted, have a low quartz content (32.6% average, ranging from 24.4% to 42.6%), and a moderately high feldspar (12.6% average, ranging from 6.8% to 23.0%) and volcanic rock fragment content (17.8% average, ranging from 7.4% to 29.8%). The correlation coefficients between samples range from 0.2683 to 0.9098, whereas the correlation coefficient between this division and the second division is 0.1558.

The second division (Groups II, III and IV) consists of 37 samples from the Tutoop Sandstone and Ketungau Formation. Correlation coefficients between samples range from 0.2638 to 0.9221. This division is characterized by a higher quartz content (55.4% average, ranging from 42.0% to 70.6%), lower feldspar content (4.6% average, ranging from 0.6% to 9.4%) and lower volcanic rock fragment content (11.1% average, ranging from 2.8% to 19.0%) than the first division. This second division can be divided into three groups (Group II, III and IV) as shown in Figure 4.68.
Group II consists of fourteen samples from the Tutoop Sandstone with the correlation coefficients between samples ranging from 0.5779 to 0.9221. The correlation coefficient between this group and group III is 0.5347. Group II represents the typical coarser samples from the Tutoop Sandstone with their characteristically large maximum grain size (0.72 mm average, ranging from 0.4 mm to 1.2 mm) and large average grain size (0.36 mm average, ranging from 0.2 mm to 0.6 mm).

Group III consists of 11 samples which are dominated by sandstone from the Ketungau Formation. The correlation coefficients between samples range from 0.5063 to 0.8920, and the correlation coefficient between this group and Group IV is 0.3718. This group represents the Ketungau Formation, which is characterized by a smaller maximum grain size (0.44 mm average, ranging from 0.3 mm to 0.4 mm) and smaller average grain size (0.21 average, ranging from 0.2 mm to 0.30 mm) than the Tutoop Sandstone.

Group IV consists of a mixture of samples from the Tutoop Sandstone (8 samples) and the Ketungau Formation (4 samples). This group represents the transition between the Tutoop Sandstone and the Ketungau Formation.

4.3.1e RELATIONSHIPS BETWEEN GROUPS AND BASINS

The relationship of sandstone samples between groups in the middle and lower Melawi Basin sequence is shown in Figure 4.69. Ninety five samples, each with 58 variables (Appendix 11), were used for Q-mode cluster analysis (Fig. 4.69) which
subdivided the samples into two major divisions.

The first division is formed by 66 samples from the Ingar Formation, Silat Shale, Payak Formation and Tebidah Formation with correlation coefficients between samples ranging from 0.0251 to 0.8697. This division is characterized by a low quartz content (33.0% average, ranging from 8.8% up to 72.0%), a low composite quartz content (6.7% average, ranging from 0.2% up to 28.0%), a high feldspar content (22.7% average, ranging from 1.7% to 51.2%) and a high volcanic rock fragment content (15.2% average, ranging from 0.5% up to 30.0%). Sandstone from the Ingar Formation forms Group Ia, whereas sandstone from the Silat Shale forms Group Ib, and sandstone from the Payak and Tebidah Formations cluster in Group Ic. Sandstone from the Silat Shale is more closely correlated with sandstone from the Payak and Tebidah Formations, having a correlation coefficients between the subgroups of 0.0972, than with the Ingar Formation. The correlation coefficient between these units and sandstone from the Ingar Formation is 0.0805.

The second division consists of 29 samples from the Dangkan and Sepauk Sandstones with correlation coefficients between samples ranging from 0.1046 to 0.7886. This division is characterized by a higher quartz content (57.2% average, ranging from 27.5% to 80.0%), a higher composite quartz content (16.3% average, ranging from 4.0% to 42.0%), a lower feldspar content (4.4% average, ranging from 0.0% to 15.0%) and a lower volcanic rock fragment content (8.8% average, ranging from 0.0% to 26.5%).
Eighty three samples, each with 58 variables (Appendix 11), were used for Q-mode cluster analysis in order to test the correlation between sandstone samples from the Melawi and Kapuas Groups (Fig. 4.70). Based on the Q-mode dendrogram (Fig. 4.70) the samples from the Melawi and Kapuas Groups can be divided into four groups.

Group I consists of 35 samples from the Payak and Tebidah Formations with correlation coefficients between samples ranging from 0.0983 to 0.8067. The correlation coefficient between Group I and Group II is 0.0423. Group II comprises 16 samples dominated by the Alat and Sepauk Sandstones, with correlation coefficients between samples ranging from 0.0836 to 0.7810. This group has a slightly negative correlation (-0.0160) with Groups III and IV. Group III consists of 11 samples dominantly from the Alat Sandstone. The correlation coefficients between samples within this group range from 0.1057 to 0.7247. This group has a negative correlation (-0.0123) with Group IV which consists of 21 samples dominantly from the Sekayam Sandstone. Correlation coefficients between samples in Group IV range from 0.0106 to 0.8340.

Based on the Q-mode dendrogram (Fig. 4.70) sandstone samples from the Payak and Tebidah Formations are weakly correlated with part of the Alat Sandstone (Group II), and negatively correlated to the remainder of the Alat Sandstone (Group III) and the Sekayam Sandstone. The Sepauk Sandstone, however, is strongly correlated to part of the Alat Sandstone (Group II) and negatively correlated to the remainder of the
Alat Sandstone (Group III) and the Sekayam Sandstone.

Three sets of Q-mode cluster analyses were used to test the correlation between sandstone samples from Melawi and Ketungau Basin sequences. Firstly between the Suwang Group (Melawi Basin) and the Merakai Group (Ketungau Basin), secondly between the Melawi Group and the Merakai Group, and finally between the Kapuas Group (Melawi Basin) and the Merakai Group.

One hundred sandstone samples from the Suwang and Merakai Groups, each with 58 variables (Appendix 11), were subjected to Q-mode cluster analysis to test for correlation between groups (Fig. 4.71). Based on this analysis the samples can be divided into two major divisions (Fig. 4.71). The first division consists of 44 samples dominated by sandstone from the Suwang Group. The correlation coefficients between samples range from 0.0313 to 0.8300. Within this division Group IA represents the Silat Shale and Group IB represents the Dangkan Sandstone. The first division has a negative correlation (-0.0331) with the second division which consists of the 56 samples from the Merakai Group. The correlation coefficients between sample within the second division range from 0.0576 to 0.8925. Within this second division the Kantu Formation is represented by Group IIA, the Tutoop Sandstone is represented by Group IIB, and the Ketungau Formation is represented by Group IIC.

In order to test the correlation between the Melawi and Merakai Groups, 100 samples from the two groups, each with 58 variables (Appendix 11), were used in Q-mode cluster analysis.
(Fig. 4.72). Based on this analysis sandstone samples from the Melawi and Merakai Groups can be clustered into two major divisions. The first division contains 60 samples from the Payak and Tebidah Formations (Melawi Group) and the Kantu Formation (Merakai Group). The correlation coefficients between samples range from 0.2130 to 0.9264, whereas the correlation coefficient between the first and the second divisions is 0.1014. The second division contains 40 samples from the Tutoop Sandstone and Ketungau Formation (Merakai Group) and 3 -samples from the Sepauk Sandstone (Melawi Group). Correlation coefficients between samples range from 0.2200 to 0.9198. Based on this analysis (Fig. 4.72) sandstone samples from the Melawi Group can be correlated with the sandstone from the Kantu Formation (Merakai Group).

One hundred sandstone samples from Kapuas and Merakai Groups, each with 58 variables (Appendix 11), were subjected to Q-mode cluster analysis to test correlation between these groups (Fig. 4.73). Based on this analysis the samples can be divided into four groups. The first group consists of 20 samples from the Alat and Tutoop Sandstones. Correlation coefficients between samples within the first group range from 0.0879 to 0.8255, whereas the correlation coefficient between the first and second groups is 0.0236. The second group consists of 20 samples from the Sekayam and Alat Sandstones with correlation coefficients ranging from 0.0615 to 0.8360. The third group consists of 26 samples dominantly from the Kantu Formation (IIIB) with correlation coefficients ranging from 0.2662 to 0.8554. Some samples from the Alat
and Tutoop Sandstones (IIIA) are also included in this group with correlation coefficients ranging from 0.2013 to 0.6324. The correlation coefficient between the first three groups and the fourth group is 0.0225. The fourth group consists of 28 samples from the Tutoop Sandstone and Ketungau Formation with correlation coefficients ranging from 0.0730 to 0.8748. This analysis indicates that the Alat Sandstone can be correlated with the Tutoop Sandstone (Group I and Subgroup IIIA).

These three sets of Q-mode analysis (Figs 4.71, 4.72 and 4.73) indicate that the most probable relationship between the Melawi and Ketungau Basin sequences are as follow: the sandstone from the Melawi Group (Melawi Basin) can be correlated with the sandstone from the Kantu Formation (Ketungau Basin) and the Alat Sandstone (Melawi Basin) can be correlated with the Tutoop Sandstone (Ketungau Basin).

4.3.2 **R-mode cluster analysis**

In the R-mode cluster analysis the data were examined using a Pearson product-moment correlation coefficient and 57 variables. Correlation between variables is an important feature leading to an understanding of the petrography of this area. The trends for each group are shown in Figures 4.74 (Suwang Group), 4.75 (Melawi Group), 4.76 (Kapuas Group) and 4.77 (Merakai Group).

Textural variables such as packing, sorting, maturity, shape and sphericity show significant linkage. In the Suwang and Kapuas Groups (Figs 4.74 and 4.76) these variables are
negatively correlated to both maximum and average grain size (-0.0352 and -0.0220). In the Melawi and Merakai Groups (Figs 4.75 and 4.77), however, the packing, sphericity, shape, sorting and maturity are poorly correlated to both maximum and average grain size (0.0054 and 0.0057). Packing, sorting and maturity are positively correlated with quartz content. It is confirmatory that the packing, sorting and maturity are better in the more rounded, finer grained sediments which also have higher quartz contents. This due to the fact that the fluvial Dangkan, Sepauk, Sekayam, Alat and Tutoop Sandstones in general are coarser grained and less sorted than the shallow marine to lacustrine sandstone in the Silat Shale, Payak, Tebidah, Kantu and Ketungau Formations.

Volcanic rock fragments and feldspar are negatively correlated with quartz and composite quartz, or in another words volcanic rock fragments and feldspar increase with decreasing quartz and composite quartz content.

Generally K-feldspar shows a strong correlation with plagioclase, but they are both poorly correlated with undifferentiated feldspar, clay minerals, kaolinite and sericite. K-feldspar and plagioclase generally occur in similar low concentrations, whereas the distribution of undifferentiated feldspar is higher due to the replacement of K-feldspar and plagioclase by clay minerals such as kaolinite and sericite.

In the Suwang Group (Fig. 4.74) quartz containing vacuoles is positively correlated with volcanic rock fragment content, but they are both negatively correlated with embayed
quartz and aphanitic lithic fragments. However, in the other groups volcanic rock fragments show a negative correlation with quartz containing vacuoles and embayments and also aphanitic rock fragments. These features may indicate that the same source supplied volcanic rock fragments and quartz containing vacuoles to the Suwang Group, but that it did not contribute significant quantities of embayed quartz or aphanitic rock fragments. Also it suggests that the source of the Suwang Group was not a major contributor during the deposition of the other groups.

Quartz with undulose extinction, quartz containing inclusions, and garnet are all positively correlated with metamorphic rock fragments, such as metaquartzite, mica and sericite schist in the Suwang Group (Fig. 4.74). This indicates that these clasts probably were derived from the same metamorphic provenance. These clasts are also positively correlated with basalt/ultramafic fragments, which suggests that the latter were also derived from the same metamorphic provenance, probably the Boyan Melange.

In all the groups, composite quartz is positively correlated with both maximum and average grain size (Figs 4.74 to 4.77). This feature indicates that the composite quartz is more abundant in the coarser grain sizes.

Limestone, micrite and sparite are highly correlated with each other in Melawi Group (Fig. 4.75). The carbonate is interpreted to be related to the shallow marine depositional environment of the Melawi Group.
4.4 IMPLICATIONS AND DISCUSSION

The influence of tectonism on the petrography of sandstone was defined by Crook (1974) and his model was supported by Schwab (1975). More recently this model has been refined by Dickinson and Suczek (1979) and Ingersoll and Suczek (1979). The latter concentrated on the composition of ancient sandstone where tectonic setting could be inferred using data other than sandstone composition. Dickinson and Valloni (1980) and Valloni and Maynard (1981), on the other hand, have examined the composition of modern sand from known plate tectonic settings. Close agreement exists between the compilations for both ancient sandstone and modern sands.

Dickinson and Suczek (1979) and Dickinson et al. (1983a) defined three provenance fields for sandstone from basins with different tectonic settings: continental block provenance, magmatic arc provenance and recycled orogen (including foreland uplift, collision orogen and accretionary prism) provenance (Figs 4.78 and 4.79). Ingersoll and Suczek (1979) presented triangular plots showing the composition of selected sand or sandstone from a variety of known tectonic settings (Fig. 4.80).

Described relationships between sandstone composition, tectonic setting and provenance (Figs 4.78, 4.79 and 4.80) were used to infer depositional settings for the sandstone from the Melawi and Ketungau Basins, as shown in Tables 4.5, 4.6 and 4.7.
4.4.1 Melawi Basin

Sandstone from the Melawi Basin ranges in composition from lithic arkose to feldspathic litharenite, sublitharenite and litharenite (classification of Folk, 1968).

The mean and standard deviation of each petrographic parameter (Figs 4.60, 4.61, 4.62, 4.63 and 4.64) show that quartz (Q), monocrystalline quartz (Qm), polycrystalline quartz (Qp), lithic (L) and lithic volcanic/metavolcanic (Lvm) detritus are all present in similar proportions in the Suwang and Melawi Groups and the Sekayam Sandstone. This uniform composition suggests that these parameters were derived from the same source area.

In the lower part of the Melawi Basin (Ingar Formation), the sandstone was deposited in a forearc area with clasts derived from a magmatic arc (Fig. 4.29; cf. Fig. 4.80), but the depositional basin was probably a long distance from the source (since the maximum and average grain sizes are 0.5 mm and 0.1 mm). The sandstone is mature with a lithic arkose to feldspathic litharenite composition (Fig. 4.26). The magmatic arc may have ranged from a dissected arc to transitional arc (Figs 4.27 and 4.28; cf. Figs 4.78 and 4.79). The QpLvmLsm data also show a very close relationship with an arc orogen source (Fig. 4.29; cf. Fig. 4.79).

In the middle of the Melawi Basin the sandstone changes in composition. In the lower Suwang Group quartz grains increase as the content of volcanic lithic fragments and feldspar decrease, and most samples have a sublitharenite composition (Fig. 4.33). The sandstone was deposited in a
basin which was fairly close to the source as indicated by the maximum and average grain sizes of 3.0 mm and 0.9 mm. In the upper part of the group, however, the sandstone composition changes to feldspatic litharenite (Fig. 4.33) and it was deposited either farther from the source area or more probably as a response to a decreased relative elevation of the source since the maximum and average grain sizes are only 0.7 mm and 0.4 mm. Both sandstone types in the Suwang Group are submature to mature.

The sandstone from the Suwang Group may represent forearc deposition with a magmatic arc source (Fig. 4.37; cf. Fig. 4.80) but the included sandstone clasts suggest derivation from a recycled orogenic provenance (Fig. 4.34; cf. Figs 4.78 and 4.79). Stratigraphically the lower part of Suwang Group might have come from a quartzose recycled area, possibly a collision suture and fold-thrust belt or subduction complex source, while the sandstone from the upper part of the group represents mixed provenance areas with some affinity to a subduction complex (Figs 4.35 and 4.36; cf. Fig. 4.79).

Sandstone forming the upper part of the Melawi Basin sequence consists of lithic arkose to feldspatic litharenite in the Melawi Group (Fig. 4.40) and sublitharenite to litharenite in the Kapuas Group (Fig. 4.47).

Sandstone samples from the Melawi Group can not be separated petrographically on the basis of formation, as can be seen in the Q-mode dendrogram (Fig. 4.66). They also have similar Lv/L and Qp/Q ratios (Fig. 4.64 and Table 4.4). The
sandstone from the Melawi Group was probably deposited in a forearc area adjacent to a magmatic arc (Fig. 4.44; cf. Fig. 4.80). Sandstone clasts probably were derived from a recycled orogenic provenance area (Fig. 4.41; cf. Figs 4.78 and 4.79), whereas based on the QmFLt diagram (Fig. 4.42; cf. Fig. 4.79) the detrital components may have been derived from a mixed provenance area. The detritus also could have been derived from a subduction complex source (Fig. 4.44; cf. Fig. 4.80).

Petrographically sandstone from the Kapuas Group can be divided into two laterally adjacent units (Fig. 4.47); sublitharenite in the eastern part of the basin (Alat Sandstone) and feldspathic litharenite to litharenite in the western part of the basin (Sekayam Sandstone). This differentiation is also clearly recognized in the Q-mode dendrogram for the Kapuas sandstone samples (Fig. 4.67). Sandstone from the Kapuas Group was deposited in either a forearc or backarc area with detritus mainly derived from a magmatic area (Fig. 4.51; cf. Fig. 4.80). Sandstone clasts in the samples probably came from a recycled orogenic provenance. The Alat Sandstone plots between quartzose recycled and transitional recycled provenance areas, whereas the Sekayam Sandstone falls into the transitional recycled provenance area (Figs 4.48 and 4.49; cf. Figs 4.78 and 4.79). The QpLvmLvm diagram (Fig. 4.50; cf. Fig. 4.80) indicates that the Sekayam Sandstone was most probably derived from an arc orogen source, whereas the Alat Sandstone is more closely related to a subduction complex source. Comparison of Figure
4.50 and Figure 4.80 shows that the Sekayam Sandstone may have been deposited in a backarc basin derived from mixed magmatic arc and rifted continental margins sources. Whereas the Alat Sandstone does not conform to any defined depositional setting.

Based on the triangular diagrams of Dickinson et al. (1983a; Fig. 4.78), Dickinson and Suczek (1979; Fig. 4.79) and Ingersoll and Suczek (1979; Fig. 4.80), most of the sandstone from the Melawi Basin was derived from a recycled orogenic provenance. The most probable recycled orogenic provenance, which occurs in the study area and may have acted as a source of the sandstone in the Melawi Basin, was the Boyan Melange and Selangkai Formation (Semitau High) to the north of the basin.

4.4.2 Ketungau Basin

Sandstone from the lower part of the Keţungau Basin consists of feldspathic litharenite. Quartz grains increase as feldspar decreases to form a predominantly sublitharenite composition in the middle part of the sequence. Feldspar and rock fragments increase slightly to produce the feldspathic litharenite composition typical of the upper part of the basin fill (Fig. 4.55). The sandstone is mainly mature. Based on the Q-mode cluster analysis (Fig. 4.68) the sandstone from the middle part of the basin has a close relationship to the sandstone in the upper part of the basin.

The sandstone in the Ketungau Basin probably represents forearc deposits, with detritus being derived from a magmatic
arc (Fig. 4.59; cf. Fig. 4.80). The sandstone may also have been derived from a transitional recycled orogen provenance for the lower and upper parts of the group, whereas the sandstone from the middle part was derived from a quartzose recycled orogenic provenance (Figs 4.56 and 57; cf. Figs 4.78 and 4.79). Comparison of Figures 4.58 and 4.79 shows that the sandstone from the Ketungau Basin plots very close to the subduction complex source, whereas comparison of Figures 4.58 and 4.80 shows that the sandstone plots at outside any defined depositional settings.

Based on these triangular diagrams, the most suitable provenance for the sandstone in the Ketungau Basin was the Lubok Antu Melange which crops out to the north of the basin.

Based on the Q-mode cluster analysis dendrogram (Fig. 4.72) the sandstone from the lower part of the Ketungau Basin (Kantu Formation) has a close petrological relationship with sandstone from the middle part of the Melawi Basin (Melawi Group). Petrographically the sandstone from middle part of the Ketungau Basin (Tutoop Sandstone) can be correlated with the sandstone from upper part of the Melawi Basin (Kapuas Group).
CHAPTER 5
DIAGENESIS

5.1 INTRODUCTION

Diagenesis is the process involving all the physical, chemical and biological changes in a sediment after deposition, during and after lithification (Larsen and Chilingar, 1979; Chilingarian, 1983). It includes compaction, solution, cementation, authigenesis, precipitation, crystallization, replacement, oxidation, reduction, leaching, hydration, polymerization and bacterial action.

Over the past few years many classifications, resulting from different approaches, have been proposed for the stages of diagenesis in sandstone. An interpretative classification, which separates diagenetic processes into early, middle and late stages, was proposed by Schmidt and McDonald (1979). As shown in Figure 5.1, they subdivided the process of diagenesis into eogenetic, mesogenetic (with 5 subdivisions) and telogenetic. Helmold and van de Kamp (1984) delineated the paragenetic sequence in terms of diagenetic events and depths of burial or time as shown in Figure 5.2. This concept was developed further by Pettijohn et al. (1987) when they proposed six stages of sandstone diagenesis on the basis of depth of burial (Fig. 5.3). Pettijohn et al. (1987) also summarized the evidence of mineralogical, textural, physical and chemical characteristics of the diagenetic
processes in sandstone as shown in Figure 5.4. Burley et al. (1987) considered the importance of liquid expulsion during diagenesis. They summarized the diagenetic processes in terms of expulsion of water from mudrock causing aqueous chemical reactions in sandstone, which, in turn, leads to a series of depth-related diagenetic stages (Fig. 5.5).

By understanding diagenesis, the provenance and depositional environment of the original sand, and the post-depositional processes, can be assessed.

5.2. METHODS

Conventional petrographic (using a polarizing microscope), scanning electron microscope and X-ray diffraction techniques were undertaken in order to study the diagenetic processes affecting sandstone units in the Melawi and Ketungau Basins.

5.2.1 Petrography

The basic technique used in the assessment of diagenetic processes and products was standard thin section petrology. The samples were all impregnated with stained (blue) araldite in order to determine the detrital grains, authigenic minerals and porosity, and also to determine the structural and temporal relationships between them.

The guides illustrated by Scholle (1979) and Adams et al. (1984) were used to determine the detrital grains,
authigenic minerals and porosity.

The terms primary and secondary porosity used in this study follow the nomenclature outlined by Choquette and Pray (1970), as later modified by Hoholick et al. (1984). Primary porosity includes all pore spaces formed before or during deposition. Secondary porosity incorporates all post-depositional pores which developed after all the primary pore spaces are filled when later processes create new openings at the sites of the earlier pores or in other parts of the rock. This term includes pores created by the dissolution of framework grains and cement, and also includes porosity in fractures and shrinkage cracks (Hoholick et al., 1984). A theory for framework grain dissolution in sandstone was proposed by Siebert et al. (1984). The genetic types of pore systems constituting secondary porosity were outlined by Choquette and Pray (1970), Schmidt and McDonald (1979; as shown in Figures 5.6 and 5.7), Pittman (1979) and Shanmugam (1985).

One hundred and thirty six thin sections of sandstone from the Melawi Basin and 69 thin sections of sandstone from the Ketungau Basin were studied in order to determine the type and sequence of diagenetic processes that affected the sandstone in these basins.

5.2.2 Scanning Electron Microscopy (SEM)

Scanning electron microscopy can greatly enhance the quality of knowledge about diagenetic processes affecting sandstone sequences. Certainly SEM reveals details and
complexities of the diagenetic phenomena that cannot possibly be determined by employing more traditional petrographic methods. SEM can be used effectively to elucidate the mineralogical changes that occur during diagenesis, and also to illustrate the three dimensional relationships between detrital grains, authigenic minerals and pore geometry.

Forty samples were selected for SEM examination and the analyses were carried out using an HITACHI-S-450 SEM at the University of Wollongong. These samples had been analyzed previously using petrographic methods and thin sections. An approximately 5x10x10 mm chip of each freshly dried sample was coated with a very thin gold layer (200 Å, using an SEM Coating Unit PS3) in order to obtain a clear image of the surface. The samples were studied at magnifications ranging from 25x to 2300x. Mineral composition was identified on the basis of comparison with the figures in Welton (1984) and was checked, where appropriate, by EDAX analysis. The SEM analyses provide clear identification of dissolution, recrystallization and authigenic minerals.

5.2.3 X-ray Diffraction (XRD)

Forty samples were selected for XRD analysis using a Phillips 1130/90 diffractometer at the University of Wollongong. Copper radiation with a graphite monochromator was used and the machine was run at 1/2° 2θ/min.
The samples analyzed included representatives of the fine-grained lithologies and the fine-grained fractions of sandstone samples from each unit. The fine-grained lithologies were analyzed after the samples had been crushed into a powder, whereas the fine-grained fraction of each sandstone sample was analyzed after it had been separated from the detrital grains by gentle disaggregation and settling in water. The clay fractions from both types of sample were mounted on ceramic discs using suction techniques.

Three analytical runs were carried out on each sample as follows. The first analysis was on the oriented untreated sample, the second was after the sample had been saturation with ethylene glycol and the third analysis followed heating the sample for 0.5 hour at 450°C. The results of these analyses are summarized in Table 5.1.

5.3 DIAGENETIC PROCESSES

Based on the petrographic, scanning electron microscopic and X-ray diffraction studies the sandstone samples from the Melawi and Ketungau Basins display a variety of diagenetic features which reflect alteration by diagenetic processes such as compaction and the formation of authigenic minerals and secondary porosity.

5.3.1 Compaction

Maxwell (1964) demonstrated that well sorted quartz sand with an initial porosity of 35-40% will be compacted
to have a porosity of only 13-20% under a load of 2.1 GPa (30,000 psi) total pressure. Such pressure exist at 9144 m depth (30,000 ft) in the Earth's crust. Porosity data from numerous sandstone sequences show that compaction in nature generally reduces porosity to 15-25% in well-sorted sandstone buried by 1525 m (5000 ft) or more. Beyond this point, cement significantly reduces the remaining porosity (Helmold and van de Kamp, 1984).

Compaction effects illustrated by sandstone samples in the Melawi and Ketungau Basins are registered by detrital mica flakes. The mica is flat to slightly bent in the least-buried rocks, such as the Sekayam and Alat Sandstones. Mica in sandstone from the Melawi Group (Fig. 5.8) and Kantu Formation is slightly bent, and the amount of bending and deformation progressively increases in the more deeply buried Suwang Group (Figs 4.19 and 4.20) and Ingar Formation.

Compaction can also be recorded by the deformation of other grains. Feldspar grains can be fractured or crushed (Fig. 4.9) and quartz grains might be strained or fractured (Figs 5.9 and 5.10) in deeply buried sequences, such as within the Dangkan Sandstone. Volcanic and sedimentary rock fragments are plastically deformed in deeply buried sandstone (Figs 5.11 and 4.22). Sutured and interpenetrated grain boundaries resulting from pressure solution (Fig. 4.6) are common in the sandstone samples from the Tebidah and Payak Formations. In some cases quartz overgrowths and cements contribute to porosity.
reduction. The combined effects of compaction, pressure solution and cementation have resulted in very tight, interlocking mosaic, grain-supported fabrics with only minimal porosity preserved. This has certainly occurred in the deeply buried Dangkan Sandstone (Fig. 4.31).

In the studied area, for a given vertical section in the Melawi Basin sequence, compaction and pressure solution are least in the Sekayam and Alat Sandstones and greatest in the Dangkan Sandstone. The same diagenetic processes are also apparent in the Ketungau Basin sequence, being least in sandstone from the Ketungau Formation and greatest in sandstone from the lower part of the Kantu Formation. However, the greatest amount of compaction and pressure solution in both Melawi and Ketungau Basin sequences was recorded from the Dangkan Sandstone.

5.3.2 **Authigenic Minerals**

Merino (1975) described mineral textures diagnostic of an authigenic origin: these include overgrowths, especially on quartz, plagioclase and K-feldspar grains; cement; euhedral crystals, mainly of quartz, albite, sphene and anatase; sutured grain boundaries; intergrowths of quartz and clay minerals with delicate fabrics; pore lining and pore filling by clay minerals; grain fractures healed in situ by quartz or albite; replacement of detrital grains by calcite and laumontite.
Authigenic minerals observed within the sandstone samples from the Melawi and Ketungau Basins include quartz, clay minerals, sphene, calcite and laumontite.

5.3.2a QUARTZ

Most secondary quartz occurs as overgrowths on detrital quartz grains (Fig. 5.12). The overgrowths range up to 0.05 mm thick and may be separated from the detrital host grain by dust seams (Fig. 4.5), rims of small hematite crystals or authigenic clay coats. In some cases host grains are differentiated by the presence of acicular inclusions or deformation lamellae which are not present in the secondary quartz.

Secondary quartz also occurs as fracture fillings in detrital quartz grains. Fractures produced in quartz grains during the early stages of compaction may be subsequently healed by secondary quartz. The fracture boundaries are often delineated by dust seams or clays similar to those associated with overgrowths. Medial sutures in the fracture fillings indicate that the quartz grew inward from the walls of the fracture and they are conclusive evidence for a post-depositional origin of the fracture.

Quartz overgrowths are present in most of the sandstone samples from both the Melawi and Ketungau Basin sequences. Quartz overgrowths progressively increase in abundance in the more deeply buried sandstone units. In the Melawi Basin sequence, the least evidence of quartz
overgrowths was found in the Sekayam and Alat Sandstones, and the greatest evidence of quartz overgrowths was found in the Dangkan Sandstone (Fig. 4.31a,b). A similar pattern is evident in the Ketungau Basin sequence where the least evidence of quartz overgrowths is in the sandstone samples from Ketungau Formation while the greatest evidence of secondary quartz is in the sandstone from lower part of the Kantu Formation. The greatest abundance of quartz overgrowths from either basin occurs in the Dangkan Sandstone (Fig. 4.31a,b).

Other occurrences of secondary quartz include small euhedral crystals of quartz growing within pore spaces (Figs 5.13 and 5.14). They occur in most sandstone samples.

5.2.2b CLAY MINERALS

Authigenic clay minerals form the main cement in many sandstone samples from both the Melawi and Ketungau Basins. The clay forms an average of 5.3% of the Melawi Basin sandstone samples but ranges from 0% to 21.2% in abundance. In the Ketungau Basin the sandstone samples contain an average of 4.6% clay and the range is from 0% to 9.8%. Two types of authigenic clay minerals are present in the Melawi and Ketungau Basins. One type occurs as an authigenic mineral lining or filling the pore spaces. The second type occurs as a replacement mineral, generally replacing feldspar and volcanic fragment grains (Fig. 4.10).
The authigenic origin of the clay is interpreted by the identification of clay rims or clay coats that form rinds around detrital sand grains (Galloway, 1974; Wilson and Pittman, 1977). It commonly forms homogenous rims of uniform distribution and thickness (up to 10 μm). Clay coated detrital grains may have formed at an early stage of diagenesis. Clay rim cements are commonly found in the sandstone samples from the upper part of both the Melawi and Ketungau Basin sequences, i.e. the Alat and Sekayam Sandstones and the sandstone from the Ketungau Formation.

Clays coatings may be replaced or shoved aside by the force of crystallization of later cements. Such circumstances are commonly found within the sandstone samples from the more deeply buried units, such as the Payak and Tebidah Formations in the Melawi Basin and the Tutoop Sandstone and Kantu Formation in the Ketungau Basin. Authigenic clay generally occurs as thin partial coatings immediately adjacent to the detrital quartz grain surfaces and represents the earliest authigenic component to form. Idiomorphic quartz overgrowths are usually the next form to develop and they are attached to quartz nuclei and tend to cover or enclose the clay coatings (Figs 4.5 and 5.14).

Clay minerals also cover some other authigenic components and must, therefore, be authigenic in origin. In some cases quartz overgrowths partly cover the clay coating, but clay coatings may have been produced more than once on some grains. This indicates that the
sandstone may have undergone more than one phase of clay forming diagenetic process, as in the Dangkan Sandstone.

Authigenic clay minerals are also present along fractures, which were produced during compaction of individual grains, as well as on the adjacent grain surfaces. This circumstance provides very useful evidence for the detection of secondary quartz which occurs as fractures fillings in detrital quartz grains.

Based on the scanning electron microscopic study and X-ray diffraction analyses, the authigenic clays which occur within the sandstone samples from the Melawi and Ketungau Basins include kaolinite, chlorite, smectite (montmorillonite) and illite.

Smectite is the most common authigenic clay within the sandstone from both the Melawi and Ketungau Basins. It appears as a cement filling pores in the sandstone and is characterized by highly crenulated, honeycombed, interlocking crystals (Fig. 5.15a,b).

Chlorite and kaolinite are the second most abundant authigenic clays found in the studied area. Chlorite generally occurs within pore spaces associated with authigenic quartz (Fig. 5.16a,b). This clay characteristically consists of numerous individual idiomorphic crystals which are plate-like in form and are attached to the detrital sand grains along their thin edges. Authigenic chlorite also forms rosette-shaped clusters associated with larger quartz crystals. Kaolinite characteristically occurs as vermicular stacks
of pseudohexagonal plates. It is generally the most readily recognizable of all the clays which grow within a pore space (Fig. 5.17). The vermicular kaolinite stacks are generally curved and intertwining with considerable intercrystalline porosity.

Illite has been detected only in sandstone from the Suwang Group where it is associated with authigenic smectite (Fig. 5.18a,b). It appears as a grain coating and a fibrous pore bridging mineral.

Authigenic clay minerals also occur as replacement minerals and, as such, they are commonly found within the sandstone from most of the units in both the Melawi and Ketungau Basins. Replacement clay mineral contents increase in the more deeply buried units.

Authigenic clay minerals may replace feldspar (Fig. 4.10) and volcanic rock fragments (Fig. 4.8) by developing along the cleavage and fracture surfaces in the grains. In the deeply buried units such as the Dangkan Sandstone, kaolinite, sericite and chlorite often partially or completely replace feldspar and volcanic rock fragments (particularly volcanic glass). Some features or structures in the original detrital grains may be preserved even after extensive replacement, e.g. faint traces of cleavage or twinning in feldspar may be preserved by the alignment of authigenic clay crystals such as sericite. Replaced feldspar laths in volcanic rock fragments, whose matrix has been replaced by
chlorite, may preserve a relict texture of the original volcanic rock.

5.3.2c PYRITE

Authigenic pyrite occurs in minor amounts in the sandstone beds from both the Melawi and Ketungau Basins. Large amounts of pyrite were only found in sandstone associated with carbonaceous beds, such as in the Silat Shale and Tebidah Formation in the Melawi Basin sequence, and in the Kantu (Fig. 4.21) and Ketungau Formations in the Ketungau Basin sequence. Merino (1975) interpreted pyrite that replaces woody material to have an early diagenetic origin, whereas Helmold and van de Kamp (1984) suggested that pyrite formed in local reducing environments associated with decaying organic matter.

5.3.2d LAUMONTITE

Laumontite only occurs in the lower part of the Melawi Basin sequence. Trace amounts of laumontite (less than 1% of the whole rock) are found within the Ingar Formation and the Dangkan Sandstone. It occurs as patches of cement interstitial to the framework grains, and also as a replacement mineral within some detrital plagioclase grains.

Laumontite cement is optically clear, with some cleavage and undulatory extinction (Fig. 4.24). It may fill pores that are partially lined by authigenic clay, thus indicating that the laumontite post-dates the genesis
of authigenic clay. The laumontite cement may be optically continuous with laumontite replacing detrital plagioclase grains, especially where the replacement is pervasive. These two modes are difficult to differentiate and probably formed at the same time.

Laumontite replacement of detrital plagioclase ranges from partial alteration of the grains to complete replacement. The replacement laumontite is usually murky brown and is associated with minor sericite and epidote inherited from the plagioclase precursor.

Helmold and van de Kamp (1984) concluded that the occurrence of laumontite is controlled by moderate burial depths and palaeotemperatures.

5.3.2e SPHENE

Numerous authigenic crystals of sphene have been recognized only in the Dangkan Sandstone during the petrographic study (Fig. 5.19). They occur as euhedral individual crystals which probably developed shortly after the pore-lining authigenic clays.

5.3.2f CALCITE

Calcite cement has been found in some sandstone samples from the Melawi Basin. The most abundant calcite content was detected in sandstone from the Ingar Formation where the average content is 10% (Appendix 11). In the other formations the average content of calcite is not more than 2%. Two generations of calcite are
distinguished based on their distribution and texture: early diagenetic localized calcite cement and late diagenetic replacement calcite. The earlier calcite generally fills pores that are lined with authigenic clays and other authigenic minerals, whereas the later calcite partially replaces some plagioclase and volcaniclastic grains.

5.3.2g FELDSPAR

Authigenic feldspar is very rare. It has been found only in a few sandstone samples from the Suwang Group and Ingar Formation. Authigenic plagioclase or K-feldspar is probably very difficult to find because most of the feldspar in the studied area has been altered to clay minerals during diagenesis and weathering.

5.3.3 Secondary Porosity

Secondary porosity created by the dissolution of feldspar and volcanic fragments occurs in sandstone throughout the Melawi and Ketungau Basins. Generally the best developed secondary porosity generated by dissolution occurs in sandstone from the Melawi Group in the Melawi Basin and from the Kantu Formation in the Ketungau Basin.

Dissolution commonly proceeds parallel to cleavage and fracture surfaces, which act as conduits for reactive fluids, resulting in a reticulate structure in partially dissolved grains (Fig. 5.20a,b). Dissolution also affects both plagioclase and K-feldspar grains producing circular
dissolution voids (Fig. 5.21a,b). Feldspar dissolution has affected both single grains, which are almost completely dissolved in some samples, and laths in rock fragments (Figs 5.22 and 5.23). The dissolution of volcanic rock fragments usually is not complete, but produces grains with distinct amounts of secondary porosity.

The dissolution of plagioclase and volcanic rock fragments is probably the result of hydration reactions, which are ubiquitous in volcanogenic terrains (Surdam and Boles, 1979).

Based on the genetic classification of secondary porosity from Schmidt and McDonald (1979; Fig. 5.6), secondary porosity which occurs in sandstone from the Melawi and Kutungau Basins includes dissolution of sedimentary material and some dissolution of authigenic cement. Based on the type of secondary porosity from Schmidt and McDonald (1979; Fig. 5.7), however, most secondary porosity in sandstone from the studied area consists of partial dissolution pores, elongate pores, corroded and honeycombed grains, and a few completely dissolved grains have resulted in oversized pores.

The dissolution of various framework grains occurred after initial compaction and has resulted in an increase in porosity for most sandstone samples.
5.4 DISCUSSION

Sandstone samples from the Melawi and Ketungau Basins display diagenetic features indicative of both mechanical and chemical diagenesis. The distribution of the features is summarized in Table 5.2.

Compaction is the main mechanical event, and began shortly after deposition with a limited amount of grain slippage and rotation. This stage of compaction is illustrated by the textures in the Sekayam and Alat Sandstones in the Melawi Basin and in the Tutoop Sandstone and Ketungau Formation in Ketungau Basin. Stronger compaction occurred in the more deeply buried sequences such as the sandstone in the Melawi Group (Fig. 5.8) in Melawi Basin, and in the Kantu Formation in the Ketungau Basin. The most intensely developed compaction was recorded in the deeply buried Dangkan Sandstone which is characterized by bent mica (Figs 4.19 and 4.20), broken feldspar grains (Fig. 4.9), strained or fractured quartz grains (Figs 5.9 and 5.10), and plastically deformed sedimentary and volcanic rock fragments (Figs 4.22 and 5.11).

Chemical diagenesis also began shortly after deposition, and has had the greatest effect on the sandstone. Authigenic clay rims and grain coatings began to form first and were followed pore lining and pore filling clay cements. This process occurred at shallow depth and was detected within the Sekayam and Alat Sandstones in Melawi Basin, and within the Tutoop
Sandstone and Ketungau Formation in the Ketungau Basin. Kaolinite pore fillings are commonly found within sandstone from the Melawi and Suwang Groups in the Melawi Basin which indicates a later deep subsurface diagenetic environment (Helmold and van de Kamp, 1984; Fig. 5.2). Dissolution of feldspar and volcanic rock fragments also began to occur at shallow depth. The relict clay rims and coats indicate that significant amounts of authigenic clay had formed before the grains were extensively dissolved. Dissolution of framework grains started at relatively shallow depths (Melawi Group and Kantu Formation) but has also been extensive at greater depths where grains were dissolved after laumontite pore filling cement had formed in the Suwang Group and the Ingar Formation. Calcite cementation was a relatively late diagenetic event that also occurred at a moderate depth of burial (Helmold and van de Kamp, 1984; Fig. 5.2) in sandstone from the Ingar Formation, Suwang Group and some from the Melawi Group.

Based on diagenetic processes (Table 5.2) as shown in the Q-mode cluster analysis dendrogram (Fig. 5.24), the sandstone samples from the Melawi and Ketungau Basins can be divided into three groups. The first group includes sandstone samples from the Kapuas Group in the Melawi Basin and the Tutoop Sandstone and Ketungau Formation in the Ketungau Basin (Group A). This group is characterized by weak compaction and the presence of clay grain coatings which indicate a mesogenetic semi-mature diagenetic stage (Schmidt and McDonald, 1979; Fig. 5.1). These features
can also be equated to the diagenetic stage 3 of Pettijohn et al. (1987) which indicates a depth of burial of about 1500 m, a temperature of about 50°C and a pressure of about 375 bar. When compared with the mudrock diagenetic stages from Burley et al. (1987; Fig. 5.5) sandstone from this group can be included in the mesogenetic mudrock stage II diagenesis with a depth of burial of about 2.4 km and a temperature in the range from 65°C to 80°C.

The second group consists of sandstone samples from the Melawi Group in the Melawi Basin and from the Kantu Formation in the Ketungau Basin (Group B). This group is characterized by the presence of common secondary porosity produced by the dissolution of feldspar and volcanic rock fragments. Based on the diagenetic cycle of Schmidt and McDonald (1979; Fig. 5.1), this group can be equated to the mesogenetic mature A stage of diagenesis. Using the classification of Pettijohn et al. (1987; Fig. 5.3) this group would be included in stage 4 which suggests a depth of burial of about 5400 m, a temperature about 180°C and a pressure of about 1350 bar. Based on the diagenetic scheme for mudrock (Burley et al., 1987; Fig. 5.5), the sandstone samples from group B would be included in the lower part of mudrock stage II, with a temperature in the range from 80°C to 95°C and the depth of burial ranging from 2.4 km up to 3 km.

The third group consists of sandstone from the Ingar Formation and Suwang Group (Group C) which is characterized by the presence of laumontite cement, an
interlocking mosaic texture owing to extreme pressure solution and quartz cement (Fig. 4.31), and carbonate cementation. Based on the diagenetic cycle of Schmidt and McDonald (1979; Fig. 5.1), these sandstone samples would represent the mesogenetic mature B stage of diagenesis, whereas they would fall into the diagenetic stage 5 of Pettijohn et al. (1987; Fig. 5.3). The latter classification suggests a depth of burial of about 9600 m, a temperature of about 320°C and a pressure of about 2400 bar. Based on the diagenetic scheme for mudrock (Burley et al., 1987; Fig. 5.5), the sandstone from the third group would be included in stage III which indicates a depth of burial ranging from 3 km to 4 km and a temperature ranging from 95°C to 120°C.

On the basis of diagenetic maturity shown by the stratigraphic sequence in the Melawi and Ketungau Basins, the Sekayam and Alat Sandstones from the Melawi Basin can be correlated to the Tutoop Sandstone from the Ketungau Basin. Sandstone from the Melawi Group (Melawi Basin) can be correlated with the sandstone from the Kantu Formation (Ketungau Basin) but the sandstone samples from the Suwang Group and Ingar Formation have no correlatives in the Ketungau Basin.

Based on the diagenetic stages defined by Pettijohn et al. (1987; Fig. 5.2) the depth of burial of the lowest part of the Melawi Basin sequence was about 9600 m, whereas based on the diagenetic scheme for mudrock devised by Burley et al. (1987; Fig. 5.5) the depth of burial
would only have been about 4000 m. Since the thickness of the Melawi Basin sequence is about 7850 m, the estimated depth of burial for the lowest part of the Melawi Basin based on the scheme from Pettijohn et al. (1987) is more realistic. However, temperatures from the Burley et al. (1987; Fig. 5.5) model are probably more reasonable. The vitrinite reflectance of the shaly coal within the Ingar Formation (0.68% $R_{\text{max}}$; see Chapter 6) was produced by a temperature of about 82°C using the Karweil diagram (from Bostick, 1973) which is very close to the temperature indicated by the diagenetic stage III of Burley et al. (1987). These features indicate that the development of the diagenetic stages in the Melawi and Ketungau Basins was controlled by a temperature of burial as proposed by Burley et al. (1987), and a depth of burial similar to the thickness proposed by Pettijohn et al. (1987).

In conclusion the diagenetic stage in the Melawi and Ketungau Basins can be divided into three stages (Table 5.2). Firstly, the early shallow burial stage which is represented by the Ketungau Formation, Tutoop Sandstone and Kapuas Group where the maximum preserved cover is 1500 m. This stage is characterized by weak compaction the presence of grain coating secondary clays. The second stage is a deeper burial stage represented by the Kantu Formation and Melawi Group where the maximum preserved cover is about 3000 m. It is characterized by the presence of common secondary porosity produced by the dissolution of the feldspar and volcanic rock fragments.
Finally, the deeper burial stage which is represented by the Suwang Group and the Ingar Formation has a maximum preserved cover of about 5850 m. It is characterized by the presence of laumontite cement, an interlocking mosaic texture owing to extensive pressure solution, and quartz and carbonate cements. These diagenetic stages may be generally representative of forearc basins.

Relationships between the diagenetic processes recorded in the sandstone units from the Melawi and Ketungau Basins are shown in the R-mode cluster analysis dendrogram (Fig. 5.25). Plagioclase overgrowths, K-feldspar overgrowths and laumontite show almost perfect correlation between each other since these diagenetic processes only occur within the sandstone from the Ingar Formation and Suwang Group (Group C in the Q-mode dendrogram, Fig. 5.24) where they are all very minor. Quartz overgrowths and compaction also show complete correlation indicating that in the studied area quartz overgrowths are well developed in the well compacted sandstone (e.g. the Dangkan Sandstone) or, in other words, quartz overgrowths and compaction both increase with increasing depth of burial. Kaolinite and calcite cement also have a very high correlation coefficient indicating that kaolinite and calcite mainly occur together within the same sandstone samples. Most of the diagenetic processes show a negative correlation with dissolution. This is simply because dissolution increases the porosity
of the sandstone, whereas the other processes cause a reduction in the porosity.

Clays, barite cement and Fe oxide coatings do not correlate clearly with the rest of the diagenetic processes.
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CHAPTER 6
ORGANIC PETROLOGY

6.1 INTRODUCTION

Twenty eight outcrop and two shallow core samples were studied to determine the macerals present in coal and the type of dispersed organic matter present in the clastic sediments. Data from 14 coal samples provided by the Directorate of Mineral Resources, Bandung, were also used.

About 10 g of each sample was mounted in an Astic resin block and polished using the procedure outlined in Figure 6.1. All the polished blocks were registered in the catalogue at the Department of Geology, University of Wollongong.

The purpose of this study was to obtain qualitative information that would lead to an understanding of the occurrence, properties and significance of the three major maceral groups in coal samples and dispersed organic matter in sedimentary rocks within the stratigraphic sequence. Standard descriptive procedures for organic petrology were introduced by the International Commission for Coal Petrology (1971, 1975), Cook (1982) and Stach et al. (1982).

A Leitz MPV 1 microphotometer, calibrated against synthetic spinel and garnet standards of 0.42%, 0.92% and 1.72% reflectance, was used to measure the vitrinite reflectance and hence determine the level of maturation or organic metamorphism. Reflectance measurements were carried out using oil immersion (n = 1.518) in plane polarized light
with a wavelength of 546 nm at a temperature of $23\pm1^\circ$ C. The galvanometer was set to give a reading of one half reflectance multiplied by 100.

Vitrinite macerals not only give the best assessment of maturation since they undergo consistent changes with rank (Smith and Cook, 1980) but also show some inherent variability in reflectance according to type (Brown et al., 1964).

The reflectance measurements were carried out using the procedure defined by Cook (1982). Measurements were taken from approximately 30 particles containing vitrinite within each sample. The results of the vitrinite measurements are summarized in Table 6.1.

A Leitz Orthoplan microscope equipped with plane polarized light and UV/violet excitation fluorescence-mode (MPV 2) was used to determine the percentage of each maceral group: liptinite, vitrinite and inertinite. The microscope used for all photographs is fitted with a Leitz Vario-Orthomat camera incorporating a 5x to 12.5x zoom.

The percentages of each maceral group were determined by point counting with 500 points per sample on those samples where adequate organic matter was present for such a procedure. Visual estimates were used where the estimated dispersed organic matter abundance was less than 10%. In the latter case the relative amounts of the maceral groups were assessed at different levels of abundance such as absent (0%), rare (<0.1%), sparse (0.1-0.5%), common (0.5-2%), abundant (2-10%), major (10-40%) and dominant (>40%).
6.2 RESULTS

The results of the vitrinite reflectance measurements are summarized in Table 6.1, whereas the results of the determination of maceral types are given in Tables 6.2 and 6.3.

6.2.1 Vitrinite Reflectance

Generally the results show that the vitrinite in the Melawi Basin sequence has a reflectance ranging from approximately 0.51% $R_{\text{vmax}}$ in the Alat Sandstone to approximately 0.74% $R_{\text{vmax}}$ in the Ingar Formation (Table 6.2). In some beds the vitrinite shows a high reflectance and ranges from 0.94% $R_{\text{vmax}}$ to 3.22% $R_{\text{vmax}}$ (Fig. 6.3). The high reflectance is caused by heating associated with the igneous Sintang Intrusion. The vitrinite reflectance value is dependent on the level of organic metamorphism. For the samples studied here it is controlled by the distance of the sample from the intrusion.

Low vitrinite reflectance values characterize the lower part of the Alat Sandstone and the upper part of the Tebidah Formation. In these units the vitrinite reflectance ranges from 0.51% $R_{\text{vmax}}$ to 0.55% $R_{\text{vmax}}$ (Table 6.2).

Coal seams which have a low vitrinite reflectance are located in the upstream portion of the Kayan River. In the field these stratigraphic units are characterized by horizontal to shallow dips, with the dip angle being less than 10°.
Higher vitrinite reflectances are found in the lower part of Tebidah Formation, Payak Formation, Silat Shale Ingar Formation and Sekayam Sandstone. The vitrinite reflectance ranges from 0.57% to 0.74% $R_{v,\text{max}}$ (Table 6.2). This vitrinite occurs in coal, shaly coal or as dispersed organic matter. In the field these formations are characterized by steeper dips (with the dip angle more than 15°; Table 6.2).

In the Ketungau Basin the vitrinite reflectance varies from 0.66% to 0.78% $R_{v,\text{max}}$ (Table 6.2). The vitrinite mainly occurs in coal or as dispersed organic matter in associated sedimentary rocks. The dip of the coal seams and other sedimentary rocks in the Ketungau Basin sequence ranges from 15° to 35°.

A few samples show a high vitrinite reflectance, such as in the Kantu Formation (83RH118) where the $R_{v,\text{max}}$ is 1.54% and in the Ketungau Formation (E 54) where the $R_{v,\text{max}}$ is 0.86% (Table 6.2). The high vitrinite reflectance is again caused by the influence of the igneous Sintang Intrusions.

The Karweil diagram (Fig. 6.2; as modified by Bostick, 1973) can be used to determine an estimate of the burial temperature attained during coalification if the period of coalification is known. In explaining the Karweil model Smith and Cook (1984) showed that the model is based on first-order reaction rates and assumes that the sediment has been subjected to its present down hole temperature for its full burial history.
Temperatures indicated by vitrinite reflectance values within the Melawi Basin were 82°C for the Ingar Formation, 80°C for the Payak Formation, 78°C for the Tebidah Formation, 82°C for the Sekayam Sandstone and 75°C for the Alat Sandstone. The uniform decrease in coalification temperature from the deepest unit (Ingar Formation) in the Melawi Basin up to the Alat Sandstone (except for the value in the Sekayam Sandstone which was probably caused by an intrusion) indicates that the geothermal gradient in this area was very low, the difference only being 7°C for a 5600 m difference in overburden. The temperatures indicated by vitrinite reflectance for the Ketungau Basin sequence were 82°C for both the Kantu and Ketungau Formations.

The geothermal gradient determined from drill hole data in the West Aceh Basin, North Sumatra, is about 2°C per 100 m (Hadiyanto, pers. comm., 1991). The West Aceh Basin is in a similar tectonic setting to the Melawi and Ketungau Basins, contains a similar thickness of sedimentary fill and would be expected to have a similar geothermal gradient. The very low apparent geothermal gradients determined from the samples from the Melawi and Ketungau Basins is attributed to the fact that the samples were taken from surface outcrops with the lower units only being sampled from the flanks of structures that are known to have grown during the deposition of the upper part of the succession. They would, therefore, have suffered less burial than the total cumulative thickness of the units in either basin. If data had been available from drill holes in the central part of the Melawi Basin a more
realistic estimate of the geothermal gradient could have been determined.

6.2.2 Maceral Type

The results of the determination of maceral types in the Melawi and Ketungau Basins are given in Tables 6.2 and 6.3. Generally the percentages of maceral groups in each sample varies with stratigraphic position and geographic locality indicating both lateral and vertical variations in types. The composition of maceral groups within coal and shaly coal in the Melawi and Ketungau Basin are shown in Figure 6.4 and the mean and standard deviation of each maceral in each basin are given in Table 6.4. Vitrinite is the most abundant maceral group within both the coal and shaly coal lithologies.

6.2.2a MELAWI BASIN

The compositions, means and standard deviations of maceral groups within coal and shaly coal in the Melawi Basin are shown in Figure 6.5 and Table 6.5.

One sample of shaly coal was analysed from the Ingar Formation. It consists of 24% vitrinite, 1.2% liptinite, 0.2% inertinite and 74.6% mineral matter (Table 6.2). The liptinite includes sparse yellow to orange fluorescing sporinite, cutinite and resinite; rare yellow to orange fluorescing liptodetrinite and suberinite; and rare green to yellow fluorescing fluorinite. The inertinite mainly consists of sclerotinite (Fig. 6.6) and inertodetrinite.
Sparse bitumen and rare oil cuts, which fluoresce green to yellow (Fig. 6.7), are found in this shaly coal. Oil drops (Fig. 6.8) are also present in some places. Mineral matter consists dominantly of clay minerals with sparse pyrite and iron oxides (Table 6.3).

In the Suwang Group dispersed organic matter was only found in upper part of the group (Silat Shale), where it is present in the siltstone and claystone beds. Unfortunately the only two suitable available samples have been influenced by the Sintang Intrusion, so the organic matter has been metamorphosed to the extent that no liptinite has been preserved. The only organic matter which is present consists of sparse vitrinite with \( R_{\text{v,max}} \) of 0.96 and 3.22% (Table 6.2) and rare inertinite.

Organic matter in the Melawi Group occurs as coal, shaly coal and dispersed organic matter in sedimentary rocks. Coal and shaly coal are found in both the Payak and Tebidah Formations.

Nine samples of coal, shaly coal and other sedimentary rocks were analysed from the Payak Formation. The proportions of vitrinite, inertinite and liptinite within coal and shaly coal in this formation are shown in Figure 6.5. The vitrinite dominantly consists of telovitrinite with some detrovitrinite and gelovitrinite. The liptinite includes sparse yellow to orange fluorescing sporinite (Fig. 6.9a,b) and liptodetrinite; sparse to common yellow to orange fluorescing cutinite (Fig. 6.9a,b) and resinite (Fig. 6.10a,b); rare yellow to orange fluorescing
suberinite (Figs 6.11a,b and 6.12a,b) and alginite (Botryococcus; Fig. 6.13a,b); and rare yellow to green fluorescing fluorinite and exsudatinite. Related organic matter found in some sedimentary rock samples from the Payak Formation includes rare phytoplankton within a claystone sample (Fig. 6.14a,b). Some of the phytoplankton resemble fragments of dinoflagellate or acritarch cysts. Mineral matter found in the organic-rich samples from the Payak Formation consists of common iron oxides and abundant pyrite.

Twelve samples of coal, shaly coal and other sedimentary rock were analysed from the Tebidah Formation. The proportions of vitrinite, inertinite and liptinite within each coal or shaly coal sample in the Tebidah Formation are shown in Figure 6.5. In the lower part of the Tebidah Formation the vitrinite dominantly consists of telovitrinite with some detrovitrinite and gelovitrinite, whereas in the upper part of the formation the content of detrovitrinite is higher than telovitrinite and gelovitrinite. The liptinite in the Tebidah Formation includes rare yellow to orange fluorescing sporinite and liptodetrinite; rare to common resinite; and rare to sparse cutinite. Rare to sparse bitumen (Fig. 6.15a,b) and rare oil drops are also found within a few samples. Mineral matter which is associated with the organic matter in the Tebidah Formation consists of common pyrite and rare iron oxides.
No sample containing organic matter was taken from the Sepauk Sandstone.

Coal seams are more common in the Alat Sandstone than in the Sekayam Sandstone of the Kapuas Group. As a consequence only one coal sample was analysed from the Sekayam Sandstone compared with seven from the Alat Sandstone. The proportions of vitrinite, inertinite and liptinite within the coal from the Sekayam and Alat Sandstones are shown in Figure 6.5.

In the coal seam from the Sekayam Sandstone the vitrinite dominantly consists of telovitrinite with rare detrovitrinite, whereas the liptinite includes rare resinite. No inertinite has been found in this coal seam. Mineral matter within this coal seam consists of only sparse pyrite.

Vitrinite within coal seams in the Alat Sandstone dominantly consists of telovitrinite (13-53.6%) and detrovitrinite (28.8-61%) with some gelovitrinite (2-7.5%). Most samples have a larger amount of detrovitrinite than telovitrinite. Liptinite in the Alat Sandstone samples consists of cutinite (1.3-10%), liptodetrinite (1.2-7%), resinite (trace-7.2%), suberinite (trace-3.2%), sporinite (trace-2.2%), bitumen (0-5.8%), alginitie (0-1.5%), fluorinite and exsudatinite (0-0.6%). Inertinite consists of semifusinite (0-1%), sclerotinite (rare-3.1%), inertodetrinite (0-2.5%) and macrinite (0-0.5%) while mineral matter includes pyrite (0-3%) and clay (0-10%).
6.2.2b KETUNGAU BASIN

Organic matter in the Ketungau Basin is present as coal seams and dispersed organic matter within associated sedimentary rocks. Coal samples were analysed only from the Ketungau Formation, whereas from the Kantu Formation all the analysed samples only contained dispersed organic matter. No organic matter or coal was found in the Tutoop Sandstone. Vitrinite in the Ketungau Basin dominantly consists of telovitrinite and detrovitrinite with minor gelovitrinite. The liptinite in the Kantu Formation includes rare to common yellow to orange fluorescing resinite; sparse sporinite (Fig. 6.16a,b) and cutinite; and rare yellow to green fluorescing fluorinite. The inertinite consists of rare sclerotinite and semifusinite. Mineral matter includes common iron oxides and abundant pyrite.

Vitrinite in the Ketungau Formation dominantly consists of telovitrinite (15-90.6%) and detrovitrinite (2-65%) with some gelovitrinite (1.6-7.6%). The liptinite includes yellow to orange fluorescing cutinite (0.8-2%), liptodetrinite (0.2-1.6%), alginite (0%-trace), resinite (trace-0.6%), sporinite (trace-0.2%), suberinite (trace-1%), bitumen (0-0.5%) and exsudatinite (0%-trace). The inertinite consists of semifusinite (trace-1%), sclerotinite (0.6-1%), inertodetrinite (0.2%) and macrinite (0%-trace). Mineral matter consists of abundant pyrite and common clay.
6.3 DEPOSITIONAL SETTING BASED ON MACERAL TYPE

The maceral composition of coal and dispersed organic matter found in a sedimentary sequence, in general, indicates the environment of deposition of the original peat or sediment. Styan and Bustin (1983) showed a comparison between Carboniferous, Tertiary and modern peat forming floras and the type of resulting coal (Table 6.6), whereas Diessel (1984) described a comparison of fossil plant types from the northern and southern hemispheres (Table 6.7). Styan and Bustin described the relationship between peat types and associated coal maceral precursors (Fig. 6.17). Diessel (1984) also described the major coal-forming facies and their characteristics in typical Gondwana peat swamps (Fig. 6.18). Rimmer and Davis (1988) observed that in the Lower Kittanning seam, western Pennsylvania, telocollinite (telovitrinite) was commonly preserved in the central part of the basin whereas high desmocollinite (detrovitrinite) was preserved towards the margins.

The proportions of vitrinite, inertinite and liptinite in the coal and dispersed organic matter within the sedimentary sequences in the Melawi and Ketungau Basins (Fig. 6.4) show a dominance of vitrinite. The microlithotype of coal and dispersed organic matter in the sedimentary rocks from the Melawi and Ketungau Basins is mainly vitrite. Based on Diessel (1984; as shown in Figure 6.18) the coal seams and dispersed organic matter in the Melawi and Ketungau Basins probably originally formed in a wet forest. The vitrinite in the Melawi and Ketungau Basins dominantly consists of
telovitrinite and detrovitrinite which indicates that the coal was probably deposited as brackish water sedge-grass peats (Styan and Bustin, 1983; Fig. 6.17) where high telovitrinite indicates the central part of the basin and high detrovitrinite indicates marginal regions.

The presence of Botryococcus (Fig. 6.13a,b) and phytoplankton (Fig. 6.14a,b) in the Payak and Tebidah Formations indicates that during the deposition of these coal macerals the Melawi Group area was influenced by open water or marine conditions.

6.4 COAL AND HYDROCARBON POTENTIAL

6.4.1 Coal Potential

Generally coal seams are found only within the Tebidah Formation and Alat Sandstone in the Melawi Basin, and in the Kantu and Ketungau Formations in the Ketungau Basin.

In the Tebidah Formation coal seams are found the upper part intercalated with mudstone and sandstone beds. The thickness of individual coal seams ranges from 15 cm to 35 cm. Two coal seams have been defined (Tebidah 1 and Tebidah 2 seams; Ilyas, 1986). These coal seams crop out in the Kayan River to the east of the Nangatebidah Village and in the Malobot River to the north of the Nangatebidah Village. In the field the coal seams have a N265°E strike with a dip of 5° to the north. Megascopically the coal seams are a brownish black colour, are compact, have an angular to
conchoidal fracture and contain 60-70% of bright coal. Petrographic characteristics of these coal seams are shown in Tables 6.2 and 6.3. Chemical analyses of the coal seams from the Tebidah Formation are shown in Table 6.8.

Based on the physical characteristics, chemical analyses and the ASTM-ASA classification, the coal seams from the Tebidah Formation belong to the high volatile bituminous C class (Ilyas, 1986).

Coal seams within the Alat Sandstone are found in the area surrounding Mt Alat (Fig. 6.19) in the southeast corner of the study area. The coal thickness varies from 15 cm up to 4.35 m. Based on field observations, two coal seams can be recognized in the Alat Sandstone, the Alat 1 and Alat 2 Seams (Ilyas, 1986). These coal seams crop out in several places, with the best exposures occurring on the northern and southern slopes in the western part of Mt Alat at Pintas Creek (Alat 1 Seam). The outcrop can be followed for about 6 km with the thickness ranging from 3.80 m up to 4.35 m. Field observations show that the coal seam has a N95°E strike with a dip of 5° to the south. The cross-section based on 6 locations is shown in Figure 6.20. Megascopically the Alat 1 Seam is a brownish black to black colour, hard and angular with conchoidal fractures, and contains 60-80% bright coal.

The Alat 2 seam crops out on the northern slope of Mt Alat and at Campa Creek south of Mt Kumbal. It has a thickness ranging from 15 cm up to 60 cm and a dip of 5-10°. Megascopically this coal seam is hard, has a shiny black
colour and contains 70-90% bright coal with some resin and pyrite.

Petrographical characteristics of the coal seams from the Alat Sandstone are shown in Tables 6.2 and 6.3, whereas the chemical analyses are given in Table 6.8. Based on these criteria and the ASTM-ASA classification the Alat 2 Seam belongs to the high volatile bituminous B to C class whereas the Alat 1 Seam is a high volatile bituminous B class coal (Ilyas, 1986).

In the upper part of the Kantu Formation, coal seams have a thickness of less than 30 cm. No detailed observations were available for the coal seams in the Kantu Formation.

In the lower part of the Ketungau Formation coal seams range from 15 cm to 1.25 m. Generally five seams have been recognized in Ketungau Formation, namely the Lesung, Tanju, Melintang, Sekalau 1 and Sekalau 2 Seams (Ilyas, 1984).

The Lesung Seam is found on the northern limb of the Ketungau Formation (Fig. 6.21). It has a 40 cm to 45 cm thickness with a dip of 10°. Megascopically this coal shows a shiny black colour, is compact and brittle with a regular fracture, and contains 70-80% bright coal.

The thickness of Tanju Seam ranges from 45 cm to 65 cm with a dip angle of 10°. The seam is present on the northern limb of the Ketungau Formation (Fig. 6.21). Megascopically this coal has a shiny black colour, is brittle and contains 60-70% bright coal.
The Melintang Seam is found on the southern limb of Ketungau Formation (Fig. 6.21). The thickness of this seam ranges from 15 cm to 1 m, with a dip of about 15-20°. Megascopically the coal seam is shiny black and brittle.

The Sekalau 2 Seam crops out on the southern limb of the Ketungau Formation (Fig. 6.21) where its thickness varies from 20 cm to 60 cm. It has a dip of about 20-35°. Megascopically this coal seam has a blackish brown colour, contains 40-80% bright coal, is brittle with an irregular fracture and contains pyrite and nodules of resin.

The Sekalau 1 Seam is also present on the southern limb of the Ketungau Basin (Fig. 6.21). Its thickness ranges from 40 cm up to 125 cm with a dip of about 25-30°. Megascopically this coal seam is blackish brown and has 30-70% bright coal.

Petrographic characteristics of these coal seams are shown in Tables 6.2 and 6.3, whereas the chemical analyses are given in Table 6.8. Based on these criteria and the ASTM-ASA classification, the coal seams in the Ketungau Formation are included in the high volatile bituminous A class for the Lesung Seam; sub-bituminous B class for the Tanju Seam; high volatile bituminous C class to sub-bituminous A class for the Melintang Seam; and sub-bituminous A class for the Sekalau 1 and Sekalau 2 Seams (Ilyas, 1984).

Good quality high volatile bituminous B to C class coal in the Melawli Basin is present in the Alat 1 and Alat 2 Seams. Field conditions indicate that the Alat 1 Seam
outcrop can be traced for 6 km, the thickness of the seam ranges 3.80 m up to 4.35 m, and the dip angle is only 5°. The thickness of overburden is only 50 cm to 10 m (Ilyas, 1986). In the Ketungau Basin the best quality coal and the best field conditions are present in the Lesung Seam. This seam consists of high volatile bituminous A class coal with an outcrop thickness of 1.0 m to 1.1 m and a dip angle of about 10-15°.

6.4.2 Hydrocarbon Potential

The critical elements required for the formation of hydrocarbon accumulations are source rocks, a suitable level of organic maturation, reservoir rocks, a seal or cap rock and a trapping mechanism. The trapping mechanism is generally controlled by the structural geology. A summary of these elements in the Melawi and Ketungau Basins is shown in Table 6.9.

6.4.2a SOURCE ROCK

In the early 1980's organic petrological and geochemical investigations proved that coal seams could provide the essential elements for liquid hydrocarbon source potential in many oilfields, especially in Australia and Southeast Asia (e.g. Thomas, 1982; Smith and Cook, 1984; Cook et al., 1985; Cook, 1987; Cook and Struckmeyer, 1986).

Organic petrographic analyses of the sedimentary sequence from the Melawi and Ketungau Basins shows that most formations, except the Dangkan and Tutoop Sandstones,
contain dispersed organic matter and coal seams. In a clastic sediment, total organic carbon content also controls the source rock potential. The Ingar, Payak and Tebidah Formations and the Silat Shale contain coal seams and sufficient dispersed organic matter to have a good source rock potential in the Melawi Basin whereas the Kantu Formation probably has a good source potential in the Ketungau Basin.

6.4.2b MATURATION

Source rocks need to be mature if they are to produce hydrocarbons and hence allow migration of hydrocarbons into suitable traps. The maturation level of organic matter in coal and sedimentary rocks is indicated by the level of coalification which can be assessed by the vitrinite reflectance. Maturation level is controlled by temperature, time and pressure. Cook (1982) also stated that both coalification and hydrocarbon generation processes are influenced by elevated temperatures acting on organic matter over geologic time.

Vitrinite reflectance values (Tables 6.1 and 6.2) indicate that several samples have achieved a level of organic metamorphism equivalent to anthracite rank of coal (>3.00% \( R_{\text{v, max}} \)). These raised levels of metamorphism are found only as localized features adjacent to bodies of the Sintang Intrusive.

The normal level of organic metamorphism (burial metamorphism) reached by the Melawi and Ketungau Basin
sequences are within the oil generation window and have mean maximum reflectances \( R_{\text{v, max}} \) of 0.52-0.79%. Microscopical data (Table 6.3) show that some of the samples contain oil cuts (Fig. 6.7) and oil drops (Fig. 6.8b) which confirms that the level of organic metamorphism or maturation is well within the zone of oil generation. Therefore, all the source rocks within the Melawi and Ketungau Basins have the potential for generating hydrocarbons.

6.4.2c RESERVOIR

Reservoir rocks include any part of sequence which has sufficient porosity and permeability to permit the accumulation of hydrocarbons under adequate trapping conditions.

No specific study has been undertaken to determine quantitatively the porosity and permeability of sandstone units in the Melawi and Ketungau Basins. Petrographic studies show that the porosity and permeability have been substantially affected by diagenetic processes, especially the presence of authigenic minerals (quartz overgrowths and clay minerals) and secondary porosity.

Based on the diagenetic studies of sandstone samples from the Melawi and Ketungau Basins (Chapter 5), the diagenetic features, which are summarized in Table 5.2, were used as the basis of a Q-mode cluster analysis dendrogram (Fig. 5.24). The sandstone samples comprising Groups A and B on the Q-mode dendrogram most probably have good reservoir
properties, whereas the samples in Group C are considered unsuitable as reservoirs.

Group A (Fig. 5.24) includes the samples from the Alat and Sekayam Sandstones in the Melawi Basin, and the Tutoop Sandstone and Ketungau Formation in the Ketungau Basin. These sandstone samples are characterized by weak compaction which indicates that primary porosity is still present. Authigenic minerals have not developed in any significant quantity. Thus these sandstone units are suitable as reservoir rocks.

Group B (Fig. 5.24) includes sandstone samples from the Sepauk Sandstone, Payak and Tebidah Formations in the Melawi Basin, and from the Kantu Formation in the Ketungau Basin. These sandstone units are characterized by the presence of secondary porosity produced by the dissolution of feldspar and volcanic rock grains. The presence of authigenic minerals, such as quartz overgrowths and clay minerals, has not significantly reduced the porosity. Therefore, these sandstone units are also suitable reservoir rocks.

Group C (Fig. 5.24) includes the sandstone samples from the Ingar Formation, Dangkan Sandstone and Silat Shale. This group is characterized by an interlocking mosaic texture produced by extreme pressure solution and cementation (Fig. 4.31). The presence of this texture indicates that the sandstones in this group should be considered unsuitable as petroleum reservoir rocks because of their very low porosity and permeability.
6.4.2d SEALS

A suitable seal is a very important element in the formation of a potential petroleum trap. Seals usually consist of sedimentary beds, which have no effective porosity or permeability (commonly mudstone or shale), that overlie the potential reservoir sandstone.

In the Melawi Basin the formations which contain mudstone beds and overlie potential reservoir sandstone include the Payak and Tebidah Formations. The Kantu and Ketungau Formations provide similar seals in the Ketungau Basin.

In the Payak Formation the few interbeds of shale or mudstone (Figs. 3.50, 3.52, 3.53 and 3.54) are of limited lateral extent. Thus these fine-grained beds are not thought to be extensive enough to act as regional seals but could provide small localized traps.

As can be seen in the litho-stratigraphic correlation of the Melawi Group (Figs 3.49, 3.50, 3.51, 3.52, 3.53 and 3.54) some mudstone beds are thicker and more laterally extensive, e.g. Shale 1 and Shale 2. Shale 1 sits on the Sandstone 1 in the Tebidah Formation whereas Shale 2 sits on the Payak Formation and Sepauk Sandstone. Thus both Shale 1 and Shale 2 have very good potential to act as good seals for hydrocarbon accumulations.

In the Kantu Formation (Fig. 3.57), the middle part of the formation consists predominantly of mudstone. This mudstone would act as a good seal for the sandstone in the
lower part of the formation and for the sandstone lenses within the mudstone.

Ketungau Formation dominantly consists of mudstone (Fig. 3.59). This formation sits on top of the Tutoop Sandstone and could act, therefore, as a good seal for potential petroleum reservoirs in the Tutoop Sandstone and in the sandstone beds within the Ketungau Formation itself.

6.4.2e STRUCTURE

One of the critical factors for hydrocarbon accumulations is a trapping mechanism which is commonly provided by the presence of a suitable structure within the sedimentary sequence. Therefore, determination of the subsurface structure of a basin is very important for accessing the hydrocarbon potential of an area.

Based on the surface observations in the Melawi Basin, anticlinal structures are developed only in the central part of the basin in the Kayan River area (Fig. 2.1). Subsurface structure can be recognized only by seismic observation. Seismic surveys have been carried out by some oil companies in the study area, but unfortunately no detailed seismic information was available for the present study.

6.4.2f POSSIBILITY OF HYDROCARBON TRAPPING

Based on the preceding data which includes source rock, maturation level, reservoir rocks, seals and structure, the possibility for the formation and accumulation of
hydrocarbons in the Melawi Basin is highest in the area around the middle course of the Kayan River.

As can be seen in Fig. 2.7 the anticlinal structure in the Kayan River area clearly shows the position of the Tebidah Formation as a possible seal, the Payak Formation as a potential reservoir while possible source rocks are the Ingar Formation, Silat Shale and the Payak Formation itself. The potential of this trapping mechanism (Fig. 2.7) has been tested by Elf Aquitaine Indonesie (B. Porthault, pers. comm., 1987) by drilling the KYN-1 hydrocarbon exploration well in the Melawi Basin. Unfortunately the data and results from this drilling exploration program are not available because it is confidential information.
CHAPTER 7
DEPOSITIONAL AND POST-DEPOSITIONAL
HISTORY AND TECTONIC SETTING
OF THE MELAWI AND KETUNGAU BASINS

7.1 PALAEOCURRENT ANALYSIS

The most common sedimentary structures used as palaeocurrent indicators in the study area are cross-stratification of both planar and trough type. Other indicators measured were ripple marks and clast imbrication. Palaeocurrent indicators were measured on the outcrop localities shown in Figs 7.1, 7.3, 7.5, 7.8 and 7.10, and also at all measured sections in the area. Due to time and cost constraints when the additional field work was undertaken, palaeocurrent measurements could only be obtained from the eastern part of the study area.

The palaeocurrent measurements were corrected for tectonic tilt using a program for analyzing directional data modified after Jones (1970). The program also produced a two dimensional vectoral analysis of the data. The results are given in Appendix 7.1 and current rose diagrams for each formation were made to show the variations in current directions (Figs 7.2, 7.4, 7.6, 7.7, 7.9 and 7.11).

In the Ingar Formation, palaeocurrent indicators were very sparse and were only recorded at one location (Fig. 7.1) which yielded 3 measurements (Fig. 7.2). These measurements show that the palaeocurrent came from a
southwest direction. This palaeocurrent direction is probably not representative of the whole Ingar Formation, but at least gives an estimate of the palaeocurrent direction at that location.

Palaeocurrent indicators found in the Dangkan Sandstone are restricted to cross-beds that were recorded from several locations. Generally they can be grouped into two areas as follows: the Nangangeri and Nangadangkan areas (Fig 7.3). Both areas have palaeocurrents derived from the north (Fig. 7.4). The current rose diagrams for both locations show polymodal palaeocurrent patterns, which suggest that the cross-beds were produced in a low sinuosity braided channel environment.

In the Silat Shale, palaeocurrent indicators were recorded in the Suwangkelik River area (Fig. 7.3), where they are found within the fine- to medium-grained sandstone interbedded in the Silat Shale. The current rose diagram shows a bimodal palaeocurrent pattern (Fig. 7.4), with the dominant direction at that location indicating derivation from a northwest direction.

Palaeocurrent directions which have been recorded from the Melawi Group cover the eastern part of the study area (Fig. 7.5), but most of them represent the Tebidah Formation (8 measurement areas). The other measurements come from two locations in the Payak Formation and one location in the Sepauk Formation. The current rose diagrams for the Melawi Group (Figs 7.5, 7.6 and 7.7) show
that most of the palaeocurrents were derived from the north, whereas palaeocurrents coming from a southern direction have only been recorded from the southern limb of the Melawi Group and represent the Sepauk Sandstone (1 location) and the Tebidah Formation (1 location).

Palaeocurrent directions from the Kapuas Group have only been recorded from the Alat Sandstone (Figs 7.8 and 7.9). The current rose diagrams show that at least part of the Alat Sandstone was derived from a southern direction. The northern part of the unit has a high palaeocurrent variance but was derived from the north. Due to a lack of time when the additional fieldwork was done, no palaeocurrent indicators were recorded from the Sekayam Sandstone. According to P. Sanyoto (pers. comm., 1990), the palaeocurrent direction for the Sekayam Sandstone was from the south.

In the Ketungau Basin sequence most of the palaeocurrent indicators were recorded from the northern limb of the sequence (Fig. 7.10), only one locality area in the Tutoop Sandstone was recorded in the southern limb of the sequence. No palaeocurrent data were recorded within the Kantu Formation in the study area. According to Tan (1979) in the continuation of the Kantu Formation in the Sarawak area (the Selantek Formation), most of the palaeoflow directions were derived from the northern to northeastern part of the area. The current rose diagrams (Figs 7.10 and 7.11) show that the palaeocurrent orientations recorded within the Tutoop Sandstone and
Ketungau Formation also indicate derivation from a northern direction.

7.2 DEPOSITIONAL ENVIRONMENT

A fundamental property of genetic stratigraphic sequences constituting a basin-fill includes bedding geometry and spatial relationships within and among lithologic units. Bedding style on both regional and local scales provides much information on depositional processes and probable depositional systems or environments. Differing depositional environments or systems are characterized by an interplay or dominance of the process agents such as gravitational potential energy, confined and unconfined flows, wave and tidal energy fluxes, and permanent and barometric currents, which produce specific erosional or depositional features (Galloway and Hobday, 1983).

7.2.1 Melawi Basin Sequence

Deposition of the Melawi Basin sequence occurred in environments ranging from continental shelf to lacustrine fluvial and floodplain depositional systems. The sequence started with the deposition of the Ingar Formation, followed sequentially by the deposition of the Suwang, Melawi and Kapuas Groups.
7.2.1a INGAR FORMATION

The lower Ingar Formation is characterized by an abundance of bioturbated mudstone and a paucity of sandstone, which indicate that the environment was distal from any shoreline, i.e. equivalent to the outer shelf mud facies of Galloway and Hobday (1983). The absence of turbidite sandstone indicates that deep-sea fans and turbidity currents had no influence in the environment (Fig. 3.8). Density stratification and slow deposition from suspension is indicated by the presence of parallel laminae (Fig. 3.8). A deep outer terrigenous shelf or upper continental slope is the most likely depositional environment (Fig. 7.12).

The presence of Dinophyseae confirm a marine depositional environment but the lack of Eocene marine calcareous microfossils is unusual, especially since most samples contain reworked calcareous Cretaceous microfossils. Preferential dissolution of aragonite could possibly account for the absence of an Eocene planktonic microfauna.

Reineck and Singh (1980) and Johnson and Baldwin (1986) suggested that the main source of shelf mud is from suspension load rivers. Biological activity recorded by bioturbation on continental shelves is inversely related to the level of physical energy (Galloway and Hobday, 1983), so the presence of bioturbation in the Silat Shale indicates low energy or slow settling of suspended sediment. Galloway and Hobday (1983) also suggested that
bioturbated, pelletiferous shelf mud occurs beneath oxidized shelf water, in contrast to the laminated organic-rich mud beneath anoxic bottom waters.

Organic-rich beds found within the Ingar Formation (e.g. sample TT 176) contain oil-cuts (Fig. 6.7) and oil-drops (Fig. 6.8). Very high organic contents are produced in upwelled waters which provide large quantities of organic material to the sea bed (Galloway and Hobday, 1983). High organic concentrations in the substrate, with carbon values of 20% or more are ideal precursors of oil-prone kerogens (Demaison and Moore, 1980). However, the organic matter in the shaley coal sample analysed from the upper part of the Ingar Formation is dominated by land-derived vitrinite and includes sporinite and cutinite.

Gravity action was important in the depositional environment of the Ingar Formation and is indicated by the presence of allochthonous limestone blocks (Fig. 3.9) and slump structures (Fig. 3.8). The latter were probably caused by earthquakes which triggered slumps on basin slopes in much the same way as density currents are generated in shallower water (Walker, 1981). The presence of limestone within the mudstone in the Ingar Formation was probably originally similar to the presence of bioclastic limestone beds which form lensoidal blankets within Oxfordian shelf mudstone in southeastern Wyoming (Bouma et al., 1982), and then gravity action occurred to break the limestone lenses into allochthonous blocks.
These Wyoming carbonate units are relatively thin (0.02-1.00 m) and discontinuous. Bouma et al. (1982) suggested that the appearance of the lensoidal limestone blankets represents the effects of sediment flushing and winnowing that took place as large storm-generated waves passed over the shelf.

The increase in abundance of sharp-based storm redeposited sandstone beds in the upper part of the Ingar Formation indicates shallower water and higher energy conditions. At least two periods of shallower water facies are indicated by the measured sections, and the thin shaley coal could represent a tidal flat facies. These shallow water facies may be related to filling of the basin, uplift along the northern margin of the basin and/or fluctuating water levels due to transgressions and regressions.

7.2.1b DANGKAN SANDSTONE

The Suwang Group unconformably overlies the Ingar Formation with a marked angular divergence, especially along the northern margin of the basin. This unconformity indicates a period of uplift and erosion of the Ingar Formation and earlier sequences to the north of the Melawi Basin and reflects an early uplift phase of the Semitau High. Deposition of the Suwang Group began with the deposition of the Dangkan Sandstone in a low-sinuosity braided fluvial system and was followed by the deposition of the Silat Shale in a lagoon or quiet shallow marine
environment. It is most probably similar to model 4 of Miall (1984; Fig. 7.13).

The Dangkan Sandstone is characterized by an abundance of coarse-grained pebbly sandstone interbedded with lenses of conglomerate in the lower and central parts of the formation. Beds range from massive to parallel laminated or cross-stratified units which, in the coarse-grained facies, form poorly defined upward fining multistory sandstone bodies. In the intervening finer sandstone facies upward fining sequences 3 m to 6 m thick are more prominent. The combination of coarse grain size, sedimentary structures and upward fining sequences indicates a high energy depositional environment which can be equated to the braided fluvial model.

The braided fluvial system model has been studied in a number of modern and ancient examples. Examples of well studied modern fluvial systems include the Brahmaputra River (Coleman, 1969) and South Saskatchewan River (Cant and Walker, 1978), while examples of ancient fluvial systems include the Devonian Brownstone of Wales (Allen, 1983) and the Triassic Hawkesbury Sandstone of New South Wales (Rust and Jones, 1987).

The sandy braided fluvial model is the transitional model between distal gravelly braided rivers and braidplains (Walker, 1981; Walker and Cant, 1984). The distal gravelly braided river model was used by Rust (1972) for the middle reaches of the Donjek River as a type area model. The sand-dominated Platte and Bijou
Creek models are equivalent to the sandy fluvial model (Walker and Cant, 1984).

Sandy braided rivers consist of braid bars and broad shallow channels with the elevated areas active only during floods. The morphological elements formed in the river are shown in Fig. 7.14 (Walker, 1981; Cant and Walker, 1978; Walker and Cant, 1984). In the South Saskatchewan River in Canada (Cant, 1982), channels average 3 m to 5 m in depth below the level of nearby sand flats and they are floored by sinuous-crested dunes which build up at flood stage and deposit sets of sandy cross-beds up to 1 m in thickness. Mid- and side-channel bar deposits have planar cross-beds that average about 1 m in thickness and range up to 3 m thick. The velocity in a braided river is generally greater than 2 m/s during flood stage (Miall, 1977).

Three main bar types found in braided rivers include: longitudinal bars, comprising crudely bedded gravel sheets; transverse to linguoid bars, consisting of sand or gravel and formed by downstream avalanche-face progradation; and complex bars formed by bedform coalescence in areas of relatively low fluvial energy (Miall, 1977). Galloway and Hobday (1983) suggested that the three main bar types which occur in braided river systems are longitudinal, transverse and lateral bars.

In the Dangkan Sandstone the vertical transition within the fining upward sequences from scour marked, cross-bedded sanistone to ripple bedded sandstone (Fig.
3.21) is similar to the depositional sequence on a transverse bar produced by the low-sinuosity braided channel model (Galloway and Hobday, 1983).

The depth of flow needed to produce the cross-beds in the Dangkan Sandstone was between 1 m and 2.5 m and the maximum depth was 4 m (Table 3.1). This flow depth is similar to the average channel depth in the South Saskatchewan River (Cant, 1982). The calculated flow velocity was 2.1 m/sec, which is also similar to other braided rivers described by Miall (1977). The presence of upper flow regime plane beds with primary current lineation in the finer sandstone confirms the occurrence of shallow high velocity flows.

The lateral variation in the Dangkan Sandstone (Fig. 3.46) indicates that the coarse-grained sedimentary sequence in the western part of the basin (Dangkan River area) is more proximal in character, whereas the fine- to medium-grained sequences in the eastern part of the basin (near Suwang River) are more distal. These features suggest that the palaeoflow during deposition of the Dangkan Sandstone was derived from the west or northwest which is in accordance with the determined palaeoflow directions. The vertical variation within the Dangkan Sandstone in the Nangangeri village area (Fig. 3.21) indicates that there were two major periods of rapid coarse-grained sedimentation during the deposition of the Dangkan Sandstone. In the lower part of the formation the presence of massive to thickly bedded coarse-grained
sandstone and conglomerate indicates proximal sedimentation in the form of an alluvial fan or proximal high energy braided stream system. These coarse deposits pass up into a series of well bedded fine- to medium-grained sandstone characterized by the presence of parallel laminae and cross-beds in reasonably defined fining upward sequences. These features indicate slower and a more distal style of sedimentation in a lower gradient fluvial system — possibly a low sinuosity river on a low gradient braidplain. A repetition of this sequence occurs in the upper part of the formation and indicates a return to high energy, high gradient fluvial deposition in a proximal location. This, in turn, suggests that tectonic activity (especially uplift) had rejuvenated the source area mid way through the deposition of the Dangkan Sandstone. The top part of the Dangkan Sandstone is finer grained and indicates the re-establishment of a broad low gradient floodplain.

7.2.1c SILAT SHALE

The fine-grained nature of the Silat Shale indicates that the Melawi Basin was subsiding and the floodplain facies of the upper Dangkan Sandstone was replaced by a low energy subaqueous environment. As can be seen in the litho-stratigraphic correlation (Fig. 3.47), the Silat Shale can be divided into two parts. The lower part consists dominantly of dark grey to blackish mudstone with some intercalations of fine-grained sandstone. The upper
part consists of interbedded very fine- to fine-grained sandstone and siltstone, with some beds of medium-grained sandstone. Sedimentary structures such as ripple marks, small-scale festoon cross-beds, planar cross-beds, graded laminae, slumped lamination, tool marks and load casts typify the upper part of the formation.

Icke and Martin (1905) described a brackish water fauna from near the base of the formation, and the presence of ostracod beds (Williams and Heryanto, 1986), although rare, indicate fresh, brackish water or marine conditions. The abundance of black shale and carbonaceous siltstone suggests that the sequence was deposited under reducing conditions. The juxtaposition of shallow water, possible fresh water, reducing environments and the dominance of non-fossiliferous mudstone (Fig. 3.26) could be explained by a lacustrine model. However, the extent of the Silat Shale in the study area is at least 170 km long and 75 km wide (Fig. 2.1, Appendix 1) and the forearc tectonic setting for the Melawi Basin (Tables 4.6, 4.7) means that an extensive lacustrine environment of these dimensions is most unlikely. No modern examples of very large forearc basin lakes have been recorded, for example in the Mentawai Trough or other places with similar tectonic settings. The combination of sedimentary features in the lower Silat Shale suggests that this part of the formation was deposited in a brackish water lagoon or lower floodplain environment. Rare radiolarians (possibly reworked) suggest a marine influence in the
upper part of this interval although no calcareous Eocene microfossils have been recorded in any of the analysed samples.

The upper part of the formation has been removed by erosion from the western part of the basin. Sedimentary structures which represent deposits from turbidity current and fluidized sediment gravity flows (Middleton and Hampton, 1976; Fig. 3.24) are present in the central and upper part of the Silat Shale. The intercalated beds of turbidite sandstone indicate that periodic mass flows were a characteristic feature of the depositional environment. They include graded beds, graded laminae, tool marks, flute marks, load casts, parallel lamination, convolute lamination, and flame and slump structures within the very fine-grained sandstone. These features indicate that at least the central part of the Silat Shale was deposited in a deeper water (trough) environment, although the actual depth cannot be determined.

In some areas, especially towards the top of the formation larger sets of planar cross-bedded medium-grained sandstone (10-15 cm thick) are associated with symmetrical ripple marks and small scale festoon and planar cross-bedding. The cross-beds indicate water depths of up to 14 m with velocities probably less than 1.2 m/s. These structures were produced by moderate strength traction currents and wave action, and they indicate a shallower water marginal marine environment (Fig. 7.15).
The presence of sharp-based medium-grained sandstone beds, ranging in thickness from 50 cm to 100 cm (Fig. 3.26), probably represent storm redeposited sand, and indicate an influx of high energy either from storm waves or flooding in a source river. The vertical increase in abundance of medium-grained sandstone beds at location 87RH100 (Fig. 3.26) indicates that the energy level in the environment increased probably as a result of a regression caused by delta progradation. Exposure during the regression is possibly represented by the conglomeratic bed at location 87RH102 (Fig. 3.26). The upper part of the formation consists of a slightly deeper water succession and records the succeeding transgression. The changing water depths and sedimentary structures are very similar to the regression-transgression processes that affected the Cretaceous Cardium Sandstone in Canada (Walker and Eyles, 1988, 1991).

The presence of a moderately thick bed of tuff in the upper part of the Silat Shale indicates active felsic volcanism in the vicinity of the basin.

Based on the preserved sedimentary features in the Silat Shale, deposition began in a lagoon or lower delta plain subject to brackish water or tidal influences. Sea level rose (transgression) during mid-Silat deposition forming a trough with unstable margins. A regression and transgression in the upper part of the formation are recorded by the coarsening upward sedimentary sequences overlain by further marine deposits.
7.2.1d SEPAUK SANDSTONE

Not many data have been recorded from the Sepauk Sandstone due to limited time and access when the later field work was done. The sandstone forms a distinctive wedge-shaped deposit which is thickest near its source in the south-central part of the basin and becomes thinner to the north, where it grades into the Payak Formation, and to the east where it interdigitates with the Tebidah Formation.

Generally the Sepauk Sandstone starts with a thin conglomerate or breccia lying directly on the granitic basement. The overlying sandstone in the lower part of the formation is medium- to coarse-grained with a few pebbly sandstone lenses. The base of each fining upwards sequence is marked by a scoured erosional surface which cuts down into the finer underlying sandstone beds. Medium-scale planar tabular and trough cross-beds are common in the sandstone beds with scour surfaces between them. Flow depths for the generation of the cross-beds were probably in the range from 0.5 m to 4.4 m and velocities were probably less than 1 m/s (Table 3.1). This succession is interpreted to represent fluvial channel deposits derived from a southerly source region. The dominance of sandstone in the sequence suggests a low sinuosity or braided stream depositional environment with the main bedforms being dunes to small sand bars (wave lengths up to 40 m).
In the upper part of the Sepauk Sandstone sedimentary structures such as massive and planar beds, cross-beds and ripple marks are common, as is the presence of the mudstone interbeds. Such deposits may represent nearshore marine sand with the mudstone beds accumulating in lower energy conditions just below fairweather wave base.

A similar change in the depositional environment from fluvial channel into marine sand was recorded from the middle Atlantic shelf of USA by Bouma et al. (1982). The sandy ridges on the Atlantic shelf formed from diverse origins such as lower shoreface, inlets, capes and estuary mouths. In the Sepauk Sandstone the channel deposits may represent estuary mouth deposits which are overlain by and associated with marine sand in the upper part of the formation.

7.2.1e PAYAK FORMATION

The Payak Formation occurs as a southward thinning wedge of sandstone along the northern margin of the Melawi Basin. It grades into the Sepauk Sandstone and lower tebidah Formation. Dominant sedimentary structures present within the Payak Formation include parallel lamination and medium scale (mainly 10-15 cm) trough and planar cross-beds. The formation becomes finer grained towards the top and to the southeast.

The conglomeratic sandstone at the base of the formation and the presence of sequences of medium-grained sandstone showing upward fining and planar cross-beds
(ranging from 20 cm up to 100 cm in thickness; Fig. 3.27, location 87RH61), suggests that the lower part of the Payak Formation represents a series of fluvial or tidal channel deposits or moderate energy nearshore marine sands.

In the central and upper parts of the formation the sandstone beds are usually 10-30 cm thick and they alternate with parallel and ripple laminated grey mudstone and siltstone. Some of the sandstone beds contain groove casts on the base and ripple bedded tops. Small bivalves are found on the top of some siltstone beds and rare non-diagnostic foraminifers have been recorded (in addition to reworked Cretaceous to Paleocene microfossils) indicating marine deposition. Some carbonaceous shale, which contains alginite macerals (Botryococcus) and phytoplankton is found within the Payak Formation and indicates an open water lacustrine or marginal marine environment (Styan and Bustin, 1983; Diessel, 1984) as shown in Figs 6.17 and 6.18. The laminated mudstone interbedded in the well bedded sandstone at location 87RH58 (Fig. 3.27) contains common carbonaceous material but lacks bioturbation. It was probably deposited in an oxygen depleted lacustrine or lagoonal environment with the vitrinite-rich organic matter representing washed in terrestrial plant debris.

Shallow marine sediments have been studied by Emery (1968) and Swift et al. (1971) who emphasized the different types of shelf current, identifying intruding ocean currents, tidal currents, meteorological (storm)
currents and density currents as the four main types. Walker (1984) described a model for storm dominated shelf processes under the general headings of wind-forced currents, relaxation (storm-surge-ebb) currents and turbidity currents. In the simple case (Fig. 7.16), the wind forces water onshore, creating an elevation of the water surface and hence a seaward pressure gradient. Swift et al. (1979) and Swift and Field (1981) have shown near bottom velocities up to about 60 cm/s on the Atlantic shelf at depths of 10-20 m. Catastrophic storm flows have been monitored in the Gulf of Mexico (Forristall et al., 1977) where longshore flows reached nearly 2 m/s and seaward-directed flows were between 50 and 75 cm/s. A conceptual diagram relating major storms and their currents is shown in Fig. 7.17.

Flow velocities required to produce the cross-beds in the central part of the Payak Formation ranged from 0.44 up to 0.89 m/s; similar to the flow velocities which occur on the Atlantic shelf (Swift et al., 1979; Swift and Field, 1981) and the Gulf of Mexico (Forristall et al., 1977).

The combination of sedimentary structures, lithologies, fossils and organic material indicate that the environment of deposition for the Payak Formation fluctuated from fluvial and lacustrine to lagoonal and shallow marine. Lateral facies changes are characteristic of this formation with a general trend towards more marine conditions in the central and eastern parts of the basin.
7.2.1f TEBIDAH FORMATION

The Tebidah Formation grades laterally into and overlaps the Sepauk Sandstone and Payak Formation. Lateral facies changes are common features in the Tebidah Formation and indicate fluctuating depositional environments. Shale is a minor component in the western part of the formation but increases in abundance eastwards; the opposite is true for sandstone.

Symmetrical ripple marks, parallel lamination and small cross-beds are common within the very fine-grained sandstone in the grey mudstone dominated lower part of the Tebidah Formation. Carbonaceous matter, bivalves and gastropods are also common in some mudstone beds (Fig. 3.32). Ostracods and indeterminate planktonic foraminifers are rare. These conditions suggest a brackish water or tidal flat depositional environment for the lower part of the Tebidah Formation.

Brackish water depositional environments occur on open coasts of low relief and relatively low energy which are protected from high-energy conditions. This area constitutes the tidal flats associated with estuaries, lagoons, bays and other areas lying behind barrier islands (Fig. 7.18). Conditions necessary for the formation of tidal flats include a measurable tidal range and the absence of strong wave action (Weimer et al., 1982).

Weimer et al. (1982) subdivided tidal flats into intertidal and subtidal environments which control facies distributions. The intertidal region makes up the major
areal extent of the tidal flat. The lithologies which are present throughout the tidal flat area commonly consist of alternating layers of mud and sand. Subtidal areas represent that part of tidal flat most likely to be preserved; the deposition results from lateral accretion associated with progradation of the tidal flat and point bars associated with meandering tidal channels. Both intertidal and subtidal regions are continuously affected by tidal currents as well as wind-induced wave currents. Current velocities can be highly variable in different settings and different conditions. For example, currents are considerably higher in the Bay of Fundy tidal flats where the normal tide range is greater than 10 m, compared with tidal flats on the Gulf Coast of the United States where tides are less than 0.5 m.

Longitudinal cross-beds, megaripples and ripple laminae are major bedding types in the subtidal deposits. Interbedded sand and mud resulting in lenticular and flaser bedding are common in the subtidal facies and result from fluctuations of energy. Sand is deposited during tidal- and wave-induced current flows, and mud is deposited during slack water periods. Intertidal areas are much more likely to have bioturbation preserved especially in the topographically higher part of the flat (Weimer et al., 1982).

In the lower part of the Tebidah Formation, the common ripple marks and cross-bedding represent a subtidal facies whereas the sequence containing parallel
lamination represents an intertidal facies. The disorganized accumulations of bivalves and gastropods indicates periodic winnowing and current activity.

In the central part of the Tebidah Formation planar laminated and cross-bed sets become thicker (up to 2 m) and are associated with slump structures in the medium- to coarse-grained sandstone (Fig. 3.32). Fining upward successions recognized in these sandstone sequences represent fluvial channel deposits. The fining upward sequences are up to 6 m thick and indicate maximum channel depths. Flow depths in the Tebidah Formation channels probably ranged from 0.3 m to 4 m, but some depths of up to 14 m were calculated (Table 3.1). Corresponding flow velocities were mainly less than 1 m/s and the cross-beds were formed by the migration of dunes and some larger sand bars with wave lengths of about 130 m. The flow depth for the Tebidah cross-beds is similar to the average channel depth in the South Saskatchewan braided river in central Canada, which ranged from 3 to 5 m depth (Cant, 1982), and the average flow velocity is within the range for braided rivers described by Miall (1977; i.e. velocities of up to about 2 m/s).

Symmetrical wave generated ripple marks and small scale (3-5 cm) cross-laminated beds are common in the very fine- to fine-grained sandstone associated with parallel laminated carbonaceous beds in the upper part of the formation. Gastropods and bivalves occur sporadically in this unit. In some areas the mudstone has an intertidal
character and is associated with well laminated beach sand deposits. Bioturbation, including distinct burrows, is also present in the mudstone- and siltstone-rich upper parts of the formation. It is associated with carbonaceous shale and thin coal seams. The coal consists dominantly of telovitrinite which indicates wet conditions — possibly a brackish water sedge-grass peat (Styan and Bustin, 1983). A 25 cm coal seam in the Mentatai River section overlies a soil horizon and represents subaerial accumulation. However, it is overlain by a brackish water or marine shellbed. This upper part of the Tebidah Formation was probably deposited in a lagoon or lower floodplain lake with the sandstone beds representing splay deposits and the associated purple mudstone successions representing oxidized floodplain deposits. Alternations of individual depositional environments reflect fluctuating relative sea levels. In general, the upper part of the Tebidah Formation shows a decrease in sea level and the environment changed from a tidal flat to a floodplain with a few localized fluvial channel sandstone deposits (possibly from an anastomosing river). Similar sequences have been reported from several anastomosing rivers in the Canadian Rockies (Cant, 1982), where the channel deposits consist of very coarse-grained gravels and the floodplains are composed of fine silt, clay and peat.
7.2.1g SEKAYAM SANDSTONE

The Sekayam Sandstone occurs as a blanket deposit in the western part of the Melawi Basin and it probably grades laterally into the Alat Sandstone.

Generally the sandstone is well bedded containing both flat-laminated beds and planar cross-beds in sets 24-35 cm thick. The latter commonly have a scoured base, especially if the underlying unit is mudstone. Although not many data have been recorded from the Sekayam Sandstone, the interbedded medium- to coarse-grained sandstone (which is conglomeratic in places) and green, grey and red mudstone suggests that the unit was probably deposited by mixed load fluvial channels derived from a southwestern source area (Galloway and Hobday, 1983; Fig. 7.19). Calculated flow depths (about 1.5 m) and velocities (0.5-1 m/s) are in accord with a fluvial model where the cross-beds were formed by dunes and small sand bars (up to 40 m wave length).

Rare coal beds in the Sekayam Sandstone are dominated by terrestrial organic matter, especially telovitrinite, with minor resinite and no inertinite. The macerals in the coal indicate deposition under continuously wet conditions either as a wet forest or sedge-grass peat.

Poorly defined, fining upward sequences are present in the sandstone sequence. The predominance of laminated sandstone, the absence of fossils, the presence of thin coal seams and the intercalation of reddish oxidized
mudstone suggests that the Sekayam Sandstone was deposited in a fluvial environment.

7.2.1h ALAT SANDSTONE

Sedimentary structures in the Alat Sandstone include parallel laminated sandstone, scour marks, fining upward sequences, and medium-scale trough and planar cross-beds (mainly 18-32 cm thick, maximum 150 cm). These structures are dominant within in the medium- to coarse-grained sandstone in the lower and middle part of the formation.

In the lower part of the formation coarse lag deposits, which occur in scour pockets and are associated with coarse bedload gravel and sand, were produced in areas of maximum flow velocity. Such features indicate low sinuosity channel deposits with a proximal character. The lag deposits are associated with flat and cross-bedded sandstone in poorly defined fining upward sequences. The flow depth for cross-bed deposition in the lower Alat Sandstone ranges from 0.4 m up to 4.4 m (Table 3.1). Flow velocities vary from 0.3 m/s up to 5.25 m/s (mainly less than 1.3 m/s). This flow depth is similar to the average channel depth in the South Saskatchewan braided river in central Canada which ranged from 3 to 5 m depth (Cant, 1982), whereas the flow velocity is similar to other braided rivers which were described by Miall (1977).

The upward transition from lag to trough cross-stratification and chute-and-fill in the lower and central parts of the Alat Sandstone (Fig. 3.36) is similar to the
vertical sequence in the upstream portion of a point bar which was capped by chute-channel deposits (Galloway and Hobday, 1983; Fig. 7.21). Chutes and chute bars form during flood stage as part of the river flow cuts directly across the surface of the point bar. Significant amounts of coarse bed-load sediment move out from the main channel and are funneled through one or more chutes or channels which have been scoured into the upstream end of the point bar (Fig. 7.21). The flow spreads across the bar surface and bed-load material is deposited forming a chute bar. The chute bar contains coarse lag material typically found in the main channel. Chutes consist of relatively coarse sediment deposited by flow separation at the lee side of the bar crest.

The dominant structures in the chute complex include imbricated pebble sheets, planar laminations and mud lenses in the upstream portions of the chute channel. Trough cross-stratification commonly forms in the distal and marginal portion of a chute, planar and avalanche cross-stratification forms in the chute bar. Cut-and-fill structures are commonly associated with scour around vegetation in the chute modified point bar. The result of chute development is the deposition of coarse bed-load sediment and large sedimentary structures at the top of a point bar sequence. Finer grained sediments at the top of the sequence contain abundant small-scale structures including ripple and climbing-ripple lamination, mud drapes and root traces.
The repetition of cross-stratification and chute fill structures in the lower and middle part of the Alat Sandstone (Fig. 3.36) indicates that lateral channel migration was common in this part of the sequence. The very fine- to fine-grained sandstone which contains small-scale cross-stratification associated with ripple-marks, climbing ripples and some carbonaceous parallel lamination, may represent the upper part of the point bar sequence or associated levee or proximal splay deposits (Fig. 7.21).

Sedimentary structures such as small-scale trough cross-stratification and parallel lamination become common in the upper part of the formation within the very fine- to fine-grained sandstone. These sedimentary structures indicate that the middle and upper parts of the formation may have been deposited within fluvial channels in a meandering river system (Galloway and Hobday, 1983).

Coal seams and carbonaceous shale in the Alat Sandstone are dominated by vitrinite (especially detrovitrinite) with minor cutinite and sporinite, rare alginate, and very little inertinite. These macerals indicate accumulation under predominantly wet conditions, such as a wet forest (Diessel, 1984). The presence of rare alginate suggests some standing water in the peat swamps whereas the dominance of detrovitrinite over telovitrinite and the minor inertinite content suggest slightly dryer conditions. Alternation of wet and dry
periods is probably related to seasonal climatic variation.

Meandering rivers have a great deal of topographic differentiation (Fig. 7.20) giving their deposits highly developed cyclicity (Cant, 1982). Locally meandering rivers produce diagnostic genetic subfacies including the channel floor, point bars, chutes and chute bars, abandoned channel plugs, levees and overbank floodplains and floodplain lakes (Galloway and Hobday, 1983).

Good examples of meandering systems in modern rivers are the Mississippi and Brazos Rivers (Walker and Cant, 1984). The channel floor has coarse lag deposits including a gravelly component of clastic load, together with water-logged plant material and consolidated blocks of mud eroded locally from channel walls. Above the lag deposit, sand is transported through the bedload system. The typical bedforms on the channel floor consist of sinuous-crested dunes with heights ranging from 30 cm up to 100 cm. The sedimentary structures preserved are trough cross-stratification in the shallower part of the flow, higher on the point bar, and ripples with associated cross-lamination. Both the sequences and types of sedimentary structures preserved in the middle and upper parts of the Alat Sandstone fit the described models for sandy meandering river systems.
7.2.2 Ketungau Basin Sequence

Deposition of the Ketungau Basin sequence occurred in shallow marine to fluvial systems. It began with the deposition of the Kantu Formation, followed by the Tutoop Sandstone and the Ketungau Formation.

7.2.2a KANTU FORMATION

The 4000 m thick Kantu Formation records a number of different depositional environments that interdigitate laterally. The vertical succession also shows gradational changes from marine dominated sandstone at the base to floodplain siltstone at the top.

The basal part of the Kantu Formation consists almost entirely of medium- to coarse-grained sandstone, pebbly or conglomeratic in part. It consists of a series of superimposed fining upward multistory sandstone sequences composed predominantly of massive beds. Some flat laminated and cross-bedded sets are interspersed with the massive beds. This sequence may represent a succession of low sinuosity fluvial or tidal channel deposits or a moderate energy nearshore marine unit.

The remainder of the lower Kantu Formation consists of interbedded sandstone and siltstone. The presence of echinoid fragments, marine foraminifers and coccoliths (Pemma sp.) in this part of the Kantu Formation in Sarawak (Silantek Formation; Tan, 1979), indicate deposition under shallow marine conditions. Fossils recorded in the Kalimantan outcrops include a few molluscs (including
oysters), ostracods and echinoid spines. All the fossil evidence suggests a marginal to shallow marine depositional environment.

The sandstone units in this lower part of the Kantu Formation most probably represent a shore-zone environment (Fig. 7.22). The primary facies inherent in all shore-zone systems is the shoreface while additional associated facies are the foreshore, lagoon and tidal inlets (Galloway and Hobday, 1983). Based on a range of textures, physical structures and biogenic features, the shoreface facies can be subdivided into lower, middle and upper shoreface subenvironments.

The presence of a gravelly sediment fraction such as lenses of conglomeratic coarse-grained sandstone in the lower part of the Kantu Formation indicates that this sandstone body represents an upper shoreface environment, equivalent to the upper part of the surf zone which is dominated by powerful onshore, offshore and longshore currents. The coarse-grained sandstone layers are covered by lenses of well bedded very fine- to fine-grained sandstone (Fig. 3.57) which probably represents the foreshore environment. It corresponds to the zone of wave swash which is characterized by a dominance of planar lamination. This stratigraphic sequence is similar to the Pennsylvanian Pottsville Formation in Alabama (Galloway and Hobday, 1983), where low angle planar lamination with multiple truncation planes was attributed to foreshore
deposition overlying cross-bedded upper-shoreface quartzarenite.

The middle part of formation is composed of dark grey to black shale, mudstone, siltstone, carbonaceous laminae and coal lenses, and very fine-grained sandstone beds containing cross-laminae. These features indicate that this subunit represents overbank or floodplain deposits (Fig. 3.38). The sandstone beds represent channel margin or crevasse splay deposits and localized small channel fill deposits. The channel deposits are up to 3 m thick and become finer grained upwards. Cross-beds in the channels (average 26 cm thick) indicate flow depths of up to 3.25 m (mainly 1-2 m), velocities mainly less than 1 m/s and dune wavelengths of up to 27 m. The channel sandstone deposits are commonly associated with thinly bedded sandstone lenses attributed to an overbank splay origin. The association of overbank or floodplain deposits with carbonaceous carbonaceous matter and localized small channels probably represents an anastomosing river depositional environment (Cant, 1982) or channel margin and interchannel floodbasin facies (Galloway and Hobday, 1983) associated with backswamps (Fig. 7.23).

The only ancient system interpreted as an anastomosed river deposit is part of the Upper Mannville Group (Albian) of east-central Alberta (Putnam and Oliver, 1980, 1981; cf. Wightman et al., 1981), which comprises 35 m of channel sandstone with interchannel siltstone, shale, coal
and thin sheet-like sandstone bodies which pinch out with increasing distance from the main channel. These features are also similar to some of the channel sandstone bodies in the upper part of the Kantu Formation.

Planar and trough cross-beds, scour marks and fining upward sequences are also present within the medium- to fine-grained sandstone in the upper part of the Kantu Formation. The sandstone is present at the base of fining upward sequences 5 m to 10 m thick where siltstone usually forms the upper half of each sequence. No lateral accretion deposits were recorded from this interval. Intercalated beds of mudstone, siltstone and very fine-grained sandstone represents floodplain deposits whereas the carbonaceous shale and coal seams represent backswamp deposits. The coal consists of terrestrially derived organic matter with telovitrinite dominant over detrovitrinite. Other coal macerals include resinite, sparse sporinite and cutinite and rare inertinite. This assemblage indicates accumulation in a continuously wet environment. The common occurrence of pyrite in this coal suggests a possible brackish water or marine influence. The presence of red beds towards the top of the formation indicates that oxidation and leaching occurred when the deposits were above the local water table. The upper part of the formation probably represents anastomosing to meandering fluvial channel deposits associated with thick sequences of exposed overbank floodplain deposits and soils.
7.2.2b TUTOOP SANDSTONE

Based on lithostratigraphic correlation (Fig. 3.58) the Tutoop Sandstone can usually be divided into two parts. The lower part is dominated by coarse-grained sandstone with pebbles and conglomerate lenses. The upper part of the formation consists of well-bedded medium-grained sandstone. The grain size becomes finer westward with the appearance of some very fine- to fine-grained sandstone and mudstone beds in the Ara River area. This observation correlates with palaeoflow directions that indicate the source area lay to the northeast of the Ketungau Basin (Fig. 7.10).

The Tutoop Sandstone consists of a massive to thickly bedded sandstone with abundant medium-scale cross-beds and parallel lamination (Fig. 3.40). Basal scour surfaces, slump structures, graded beds (20-40 cm thick) and fining upward sequences are also typical of the Tutoop Sandstone. Together with the lack of fossils, these structures indicate that the Tutoop Sandstone was most probably deposited in a fluvial environment.

Calculated flow depths needed to produce the cross-beded Tutoop Sandstone ranged up to 6.2 m but most estimates were less than 4 m. This flow depth is similar to the average channel depth in the South Saskatchewan River in central Canada which ranged from 3 to 5 m (Cant, 1982). The corresponding flow velocities for the Tutoop Sandstone ranged from 0.5 m/s up to 6.2 m/s with most estimates indicating a velocity of less than 2.3 m/s, i.e.
similar to other braided river systems described by Miall (1977).

Fluvial depositional systems consist of a mosaic of genetic facies. They comprise various combinations of channel fill, channel margin and floodbasin deposits. Channel facies are the most diagnostic component, and their recognition is the key to interpretation of fluvial systems (Galloway and Hobday, 1983).

Channel deposits contain most of the bed-load sediment retained within a fluvial system and they form the skeletal framework of the system, which includes both aggradational and lateral accretional depositional units. Geometry of the channel produces characteristic internal structures. Bed accretion is a characteristic of low sinuosity sandy channels whereas lateral accretion characterizes high sinuosity channels.

Basal scours which represent the erosional channel floor (Fig. 3.39) are common in the Tutoop Sandstone. They are overlain by the coarsest material transported by the river and represent the area of maximum velocity and turbulence. Migrating subaqueous dunes cover the active channel floor in the Tutoop Sandstone and are represented by medium-scale cross-bedding. Fining upward cycles that begin with coarse channel lag deposits are also common in the Tutoop Sandstone (Fig. 3.40) and are generally less than 6 m thick. Some sandstone sequences consist of multistory channel base sandstone deposits. The presence of slump structures and overturned drag folds is also
typical of braided stream deposits (e.g. Rust and Jones, 1987). Based on these features, the Tutoop Sandstone was most probably deposited on a braidplain crossed by low sinuosity fluvial channels.

7.2.2c KETUNGAU FORMATION

The Ketungau Formation exhibits considerable lateral variation in facies with the highest proportion of sandstone in the Ara River area and the proportion of mudstone increasing both to the east and west (Fig. 3.59). The proportion of sandstone also decreases upwards through the formation.

Sedimentary structures are rare in the Ketungau Formation. They include a few planar cross-beds with an average thickness of about 0.22 m within medium-grained sandstone in the lower part of the formation. The cross-beds indicate flow depths of 0.4 m to 2.8 m, flow velocities mainly less than 1 m/s (but up to 4.2 m/s) and dune wavelengths of up to 25 m. These parameters are typical of fluvial channel deposition which is in accord with the coarse-grained pebbly nature of some of the basal lag deposits overlying scoured erosion surfaces. Parallel and carbonaceous laminae, thin coal seams and a few very fine-grained sandstone beds are present in the associated siltstone and mudstone. These sedimentary structures suggest that the intercalated mudstone, siltstone and very fine-grained sandstone represent floodplain or overbank deposits. The coal seams are largely composed of
detrovitrinite and less abundant telovitrinite, together with minor resinite, cutinite and sporinite and rare alginite. The coal macerals indicate a wet depositional environment with possible dryer periods represented by the abundant detrovitrinite. Hence, the carbonaceous material indicates a high water table during accumulation (backswamp) but the common associated pyrite suggests a brackish water or marginal marine influence.

Very fine-grained to rare medium-grained sandstone beds, ranging in thickness from 10 cm to 50 cm, are intercalated with the siltstone and mudstone. Some of the sandstone beds are slump folded, others show parallel lamination or planar cross-lamination (Fig. 3.42). The medium-grained sandstone containing only planar cross-beds most probably represents tidal flat sandstone or shallow subtidal deposits (Weimer et al., 1982). The presence of gastropods, bivalves and rare echinoids and foraminifers suggests that part of the formation was deposited in a marginal to shallow marine environment. Based on this evidence most of the Ketungau Formation was deposited in a floodplain environment with periodic shallow marine incursions due to fluctuating relative sea levels.

The upper part of the Ketungau Formation in the Ara River section is dominated by fine- to coarse-grained and pebbly sandstone. It represents a sequence of fluvial channel deposits either in a main trunk channel or as a fan-like body resulting from a local uplift of the source
area. These sandstone units may have acted as source of some intercalation sandstone within the mudstone.

7.3 PROVENANCE

In general, sandstone composition is controlled by the following factors: provenance, transportation, depositional environment and diagenesis (Suttner, 1974). The sandstone composition of the Melawi and Ketungau Basins is primarily controlled by provenance, with hydraulic sorting and diagenetic alteration only having minor influences. The interpreted tectonic setting and provenance of sandstone in the Melawi and Ketungau Basins are shown in Tables 4.5, 4.6 and 4.7.

7.3.1 Melawi Basin

Based on the 136 samples, the sandstone from the Melawi Basin ranges in composition from lithic arkose to feldspathic litharenite, sublitharenite and litharenite (classification of Folk, 1968).

Provenance and tectonic setting of the sandstone from the Melawi Basin can be determined using triangular diagrams from Dickinson et al. (1983a: Fig. 4.78), Dickinson and Suczek (1979; Fig. 4.79) and Ingersoll and Suczek (1979; Fig. 4.80), as shown in Tables 4.5, 4.6 and 4.7. Based on the QtFL triangular diagram from Dickinson et al. (1983a: Fig. 4.78) and Dickinson and Suczek (1979), most of the sandstone from the Melawi Basin was derived
from a recycled orogenic provenance, except for the sandstone from the Ingar Formation which was derived from a magmatic arc provenance ranging from a dissected arc to a transitional arc (Tables 4.5 and 4.6).

Orogenic recycling occurs in several tectonic settings where stratified rocks are deformed, uplifted and eroded. Dickinson and Suczek (1979) and Dickinson (1985) described orogenic recycling from subduction complexes, backarc thrust belts and suture belts.

Based on the QmFLt triangular diagram, the Dangkan Sandstone falls into the quartzose recycled orogen provenance, the Alat Sandstone falls into the quartzose to transitional recycled orogen provenances and Sekayam Sandstone falls into the transitional recycled orogen provenance. Sandstone from the Silat Shale, Sepauk Sandstone, Payak and Tebidah Formations falls into a mixed provenance area, whereas sandstone from the Ingar Formation falls into a dissected to transitional magmatic arc provenance (Tables 4.5 and 4.6).

The QpLvLs triangular diagram from Dickinson and Suczek (1979; Fig. 4.79) shows that most of the sandstone from the Melawi Basin was derived from a subduction complex area, but the sandstone from the Ingar Formation and Sekayam Sandstone were derived from arc orogen provenance area. The LmLvLs triangular diagram (Ingersoll and Suczek, 1979; Fig. 4.80) of sandstone from the Melawi Basin shows that the lithic fragments were derived from a magmatic arc and deposited in a forearc basin.
The relationships between mean values of each petrographic parameter from each formation are shown in Figs 4.60, 4.61, 4.62, 4.63 and 4.64. These diagrams show that quartz (Q), monocrystalline quartz (Qm), polycrystalline quartz (Qp), lithic (L) and lithic volcanic and metavolcanic (Lvm) detritus are all present in similar proportions in the Dangkan Sandstone, Silat Shale, Sepauk Sandstone, Payak and Tebidah Formations and the Sekayam Sandstone. These features indicate that the Q, Qm, Qp, L and Lvm components in these formations were probably derived from the same source area. The palaeocurrent direction for these formations (Figs 7.3 and 7.5) generally came from the north, except for the Sepauk and Sekayam Sandstones which were derived from the south. The provenance for the sandstone beds in these formations is a recycled orogenic provenance for Dangkan and Sekayam Sandstones, and orogenic recycled to mixed provenance areas for Silat Shale, Sepauk Sandstone, Payak and Tebidah Formations (Tables 4.5 and 4.6).

Based on these features, most detritus in the sandstone from these formations was derived from recycled orogenic to mixed provenance areas to the north of the basin. The most suitable provenance area for these formations is the Boyan Melange which is located on the northern margin of the basin.

Some of the monocrystalline quartz and the rock fragments which show micrographic structures (Fig. 4.14) were mainly derived from the Permian and Mesozoic granitic
rocks which crop out in the Sanggau Quadrangle northwest of the basin (Northwest Domain of Williams et al., 1988), in the Semitau High as a block in the Boyan Melange to the north of the basin (Williams and Heryanto, 1985) and in the Putusibau Quadrangle northeast of the basin (Pieters and Supriatna, 1989).

In the Dangkan Sandstone a few monocrystalline quartz grains show embayments (Fig. 4.2) and abundant vacuoles (Fig. 4.3) which indicate that these grains were derived from extrusive igneous rocks and hydrothermal-vein sources respectively. Volcanic rock fragments are also found within most of the sandstone from the Melawi Basin (Figs 4.11, 4.12 and 4.13). Extrusive igneous rocks which crop out north of the basin and could have acted as a source of the sandstone grains are the Betung Volcanics (Cretaceous age) and the Piyabung Volcanics (Eocene age).

Sedimentary rock fragments such as sandstone (Fig. 4.17) and siltstone (Fig. 4.18) are also found within the sandstone from the Melawi Basin. These sedimentary rock fragments were derived from the Selangkai Formation to the north of the basin.

Quartz grains with inclusions of sillimanite (Figs 4.1 and 4.7) and polycrystalline quartz grains with foliated structures (Fig. 4.16) indicate that these grain were derived from a metamorphic source area. Metamorphic rocks are found as blocks within the Boyan Melange (Williams and Heryanto, 1985; Williams et al., 1986). The presence of ultramafic rock fragments within the Dangkan
Sandstone (Fig. 4.15) indicates that these grains were also derived from ultramafic blocks in the subduction complex (Boyan Melange). Based on the QpLvLs triangular diagram of Dickinson and Suczek (1979; Fig.4.79) most of the Melawi Basin sequence was derived from a subduction complex, except the sandstone in the Ingar Formation and the Sekayam Sandstone which was derived from a magmatic arc (Table 4.6).

The most suitable tectonic setting for orogenic recycled detritus in the study area is the subduction complex represented by the Late Cretaceous Boyan Melange composed of a sheared matrix and blocks of chert, sedimentary, metamorphic, ultramafic, granitic and volcanic rocks (Williams and Heryanto, 1985; Williams et al., 1986). The melange belt, together with the Selangkai Formation, was tectonically uplifted to form a structural high (Semitau High) along the former trench-slope break between the trench axis and forearc basin within an arc-trench system. Sediment derived from this tectonic ridge was transported toward the arc into the forearc basin. The provenance and tectonic setting of sandstone in the Melawi Basin, which was derived from recycled orogenic provenance, is similar to that of the mid-Cretaceous Virginian Ridge Formation, Methow Pasayten sequence, Washington, and the Late Jurassic epiclastic units in the Sierra Nevada foothill belt, California (Dickinson and Suczek, 1979).
Sandstone from the Ingar Formation has a lithic arkose to feldspathic litharenite composition (Fig. 4.26) and from the Sekayam Sandstone has feldspathic litharenite to litharenite composition (Fig. 4.47) with lithic volcanic fragments dominant (Fig. 4.51). Large amounts of feldspar and lithic volcanic and metavolcanic (Lvm) fragments within the sandstone in the Ingar Formation (Figs 4.26, 4.27, 4.60, 4.61 and 4.62) and the Sekayam Sandstone (Figs 4.47, 4.48, 4.60, 4.61 and 4.62) indicate that these sandstone units were derived from a magmatic arc or arc orogen provenance ranging from dissected arc to transitional arc for the Ingar Formation and arc orogen for the Sekayam Sandstone (Tables 4.5 and 4.6).

Detritus eroded from arc orogens forms a spectrum of sand types ranging from lithic-rich volcaniclastic debris to more quartzo-feldspathic detritus of largely plutonic origin (Dickinson and Suczek, 1979). An arc with nearly continuous volcanic cover is known as an undissected arc whereas an arc where cogenetic plutons are widely exposed is known as a dissected arc.

Palaeocurrent analysis of the Ingar Formation (Fig. 7.1) and Sekayam Sandstone (Sanyoto, pers. comm., 1991) indicate that the sandstone detritus was derived from southwest of the basin.

The Schwaner Mountains, located south of the Melawi Basin, comprise tonalite, granodiorite and granitic rocks associated with regional metamorphic rocks. Intermediate to basic volcanic rocks are also present (Ammirudin and
Trail, 1987). These plutonic and associated volcanic rocks are the most suitable source for the sandstone in the Ingar Formation and Sekayam Sandstone. The provenance and tectonic setting of the sandstone from the Ingar Formation and Sekayam sandstone are similar to that of the Cenozoic Gray's Harbor-Chehalis Basins, Washington (Galloway, 1974) and the Early Cretaceous Lodoga, Great Valley sequence, northern California (Dickinson and Rich, 1972).

Cawood (1983) described how forearc petrofacies generally reflect the igneous and morphologic evolution of the arc, where eruptive rejuvenation of the surficial volcanic edifice counteracts uplift and erosion. The resultant detritus may be deposited across the forearc basin over a long period of time. Volcaniclastic sequences commonly become more quartzose, arkosic and less lithic upsection because the normal evolution of many magmatic arcs leads to eruption of more felsic material and the tendency for uplift and erosion to expose the more plutonic roots of an arc as time passes (Korsch, 1984; Dickinson and Rich, 1972).

Recycled orogenic subduction complex sources may influence forearc basin petrofacies locally (Tennyson and Cole, 1978). However in the Melawi Basin sandstone petrofacies, the recycled orogenic subduction complex source had a greater influence than the arc orogen source (Tables 4.5 and 4.6). Typical patterns for the sandstone detritus from the recycled orogenic source in the Melawi
Basin are: stable quartzose grains (Q) decrease upward with the increase in unstable lithic fragments (L) as shown in Fig. 4.60; and polycrystalline quartzose lithic fragments (Qp) decrease upward along with the decrease in lithic volcanic and metavolcanic fragments (Lvm) as shown in Fig. 4.62. Transform tectonic activity during oblique subduction may lead to especially complex patterns of forearc petrofacies in both time and space (Pacht, 1984).

7.3.2 Ketungau Basin

Fifty nine thin sections of sandstone have been studied from the Ketungau Basin sequence. The sandstone ranges from feldspathic litharenite to sublitharenite (classification of Folk, 1968; Fig. 4.55).

Based on the QtFL and QmFLt triangular diagrams of Dickinson et al. (1983a; Fig. 4.78) and the QpLvLs triangular diagram of Dickinson and Suczek (1979; Fig. 4.79), sandstone composition in the Ketungau Basin falls indicates derivation from recycled orogen and subduction complex sources as shown in Tables 4.5 and 4.6. The source ranges from quartzose recycled orogen for the Tutoop Sandstone to transitional recycled orogen for the Kantu and Ketungau Formations.

Palaeocurrent analysis indicates the source lay to the northeast (Fig. 7.10) and the most suitable provenance for the sandstone in the Ketungau Basin is the Lubok Antu Melange (Tan, 1979). This is located to the north of the basin in Sarawak (Malaysia) and trends southeastward to
the Kapuas Melange in Kalimantan (Williams et al., 1988). Tan (1979) described the Lubok Antu Melange (Early Eocene age) which is composed of a grey to dark grey, pervasively sheared and chloritised matrix and blocks of chert, sedimentary rocks, basalt, volcanic rocks and serpentinite. Pieters and Supriatna (1989) described some granitic rocks within the continuation of the Lubok Antu Melange in Kalimantan.

7.4 DIAGENETIC IMPLICATIONS

Sandstone units in the Melawi and Ketungau Basin sequences show evidence of three distinct phases of diagenesis which can be related to stratigraphic position and, hence, depth of burial. An early shallow burial diagenetic stage, represented by the Ketungau Formation, Tutoop Sandstone and Kapuas Group, is characterized by weak compaction and the presence of clay grain coatings. A second deeper burial diagenetic stage, represented by the Kantu Formation and Melawi Group, contains common secondary porosity caused by the dissolution of feldspar and volcanic fragments. The third more deeply buried diagenetic stage, represented by the Ingar Formation and Suwang Group, is characterized by the presence of laumontite cement, an interlocking mosaic texture owing to significant pressure solution, and quartz and carbonate cement.

The depth and pressure attained during burial can be estimated reasonably using the diagenetic
stages defined by Pettijohn et al. (1987). However the corresponding temperatures derived from the Pettijohn et al. model are excessive and do not relate to the coal rank or presence of hydrocarbons. The temperature attained during burial can be estimated more reliably using the diagenetic stages defined by Burley et al. (1987; see Chapters 5 and 6).

Coalification temperatures can be estimated from vitrinite reflectance (using the Karweil diagram, as modified by Bostick, 1973; see Chapter 6). In the Melawi Basin the calculated temperatures range from about 75°C in the youngest formation (Alat Sandstone) to about 82°C in the oldest formation (Ingar Formation), i.e. only about 7°C difference in temperature for about 5600 m difference in thickness. Similar data were obtained for the Kutungau Basin. These data yield geothermal gradients in the Melawi Basin of about 1°C per 800 m. When compared with geothermal gradients of about 2°C per 100 m measured from drill holes in modern forearc basins (e.g. West Aceh Basin in North Sumatra; Hadiyanto, pers. comm., 1991) the gradients in the Melawi and Ketungau Basins appear to be unrealistically low. The low gradients in these basins are probably largely due to sampling methods since the stratigraphically lower units were sampled at the surface on the margins of the basins where depth of cover due to unconformities is considerably less than in the centre of the basin.
The temperature for coalification of organic matter in the Alat Sandstone was about 75°C. Based on the general thermal gradient of 2°C per 100 m for forearc basins, the thickness of the cover above the Alat Sandstone was probably about 3750 m. This, in turn, implies that the cover thickness for the Kantu Formation and Melawi Group was about 6750 m, and for the Suwang Group and Ingar Formation was about 9350 m. Corresponding temperatures for coalification in the central parts of the basins, using the same geothermal gradient, would have been 133°C for the Kantu and Melawi Group and 183°C for the lower sequence.

Thus the diagenetic stages which occur in the Melawi and Ketungau Basins are characterized by relatively large depths of burial, high compressive loads but relatively low temperatures. The early shallow burial diagenetic stage would have formed at a depth of burial of about 3750 m, a temperature of 75°C and a pressure of about 3.75 MPa. The second stage of secondary porosity development occurs at depths of about 6750 m, temperatures of 80°C to 133°C and a pressure of about 13.5 MPa. The diagenetic stage characterized by pressure solution and cementation occurred at a depth of burial of about 9750 m, a temperature of 82°C to 183°C, and a pressure of 240 MPa. These relationships may represent typical depth-temperature profiles for the central portions of forearc basin sequences.
7.5 TECTONIC SETTING

The tectonic setting of Kalimantan and West Indonesia in general has been studied by several authors, such as Katili (1973, 1975, 1989) and Hamilton (1979, 1989).

In the Palaeozoic to Early Mesozoic, West Kalimantan had a continental basement with slate, hornfels, phyllite, quartzite, schist, gneiss, migmatite, minor metavolcanic rocks and amphibolite (Pieters and Supriatna, 1989) that was probably part of the northern margin of Gondwana (Sengor, 1987; cf Katili, 1989). Katili (1989) suggested that these rocks represent the erosional remnants of a Late Carboniferous to Early Permian subduction event in West Kalimantan that extended westwards as far as Sumatra. According to Katili's model the subduction continued or shifted slightly oceanward in Permian to Early Triassic time (Fig. 7.25).

In contrast, Hamilton (1979) suggested that the earliest subduction event recorded in West Kalimantan included the Jurassic to Early Cretaceous rocks of Kalimantan and the melange in far northwest Sarawak (Fig. 7.24). The main features in the study area related to this event are the Schwaner Mountains and the associated forearc and trench sequences to the north. The Schwaner Mountains consist of Jurassic to Cretaceous I-type granitoids and volcanic rocks (100-120 Ma) that formed from the southward dipping (present orientation) subduction zone (Katili, 1989; Hamilton, 1979). They are part of a Cretaceous and Jurassic granitic and volcanic
terrain (Hamilton, 1979; Fig. 7.24) extending northwestward along the western margin of the South China Sea.

Fault bound Permian granite bodies in the subduction complex (future Boyan Melange) probably represent microcontinents surrounded by oceanic crust that were accreted from the north (Fig. 7.26A). The oceanic crust, along with the microcontinent(s) moved towards the active Gondwana margin (Schwaner Mountains) in the Jurassic to Early Cretaceous (Katili, 1989; Hamilton, 1979) as shown in Fig. 7.26B. This tectonic setting is similar to that of the northern Sunda Arc (Mitchell, 1985), where the Mt Victoria microcontinent was accreted to the Asian crust along an east-dipping subduction zone (Fig. 7.27).

Subduction in West Kalimantan probably ceased by the mid Cretaceous and the forearc basin, inner trench slope and inactive trench on the continental margin north of the Schwaner Mountains accumulated the thick pelagic sediments and flysch represented by the Selangkai Formation (Williams and Heryanto, 1985). Sediment was probably derived from the Schwaner Zone which must have had a low to moderate relief at that time (Williams and Heryanto, 1986).

Late Cretaceous subduction produced the Boyan Melange, and the Permian granite microcontinent(s) were incorporated into this melange belt. Uplift of the Selangkai Formation and the Boyan Melange occurred during the Paleocene or Early Eocene due to motion along a set of
backthrusts which have been recorded in the Permian granite near Balaise but in the western part of the study area (Sanyoto, pers. comm., 1991). The uplifted accretionary prism was flanked to the south by a forearc basin which formed the site of the Melawi Basin (Fig. 7.26C).

Backthrusting of active accretionary wedges above forearc basins has been recognized as a common process in forearc regions (Silver and Reed, 1988). For example backthrusts occur towards the rear of subduction complexes in the Barbados ridge (Westbrook, 1982), the Mediterranean ridge (Le Pichon et al., 1982) and the Sunda arc system (Hamilton, 1979; Karig et al., 1980; Silver et al., 1983, 1985; Reed et al., 1986).

The uplifted accretionary wedge formed a local source for the sediment along the northern margin of the Melawi Basin. The common Late Cretaceous microfossils in the Ingar Formation provide proof that most of the sediment was reworked from the Selangkai Formation to the north. The presence of slump structures (see Chapter 3) and allochthonous (sandstone and limestone) blocks probably reflects rapid uplift and tectonic instability in the Semitau High on the northern margin of the basin (Fig. 7.26D).

Movement along the major backthrust at the rear of the accretionary wedge, associated with subduction, controlled the later development of the Melawi Basin sequence. Periods of deformation and uplift took place
between the Ingar Formation and the Suwang Group, between the Suwang and Melawi Groups and between the Melawi and Kapuas Groups. These deformations resulted in folding and unconformities between units within the sedimentary succession, with the greatest deformation always occurring adjacent to the Boyan Melange. The basin sequence also thickens toward the north because of uplift and overthrusting along the northern margin.

The Dangkan Sandstone contains two wedges of coarse sandstone and conglomerate indicating two phases of uplift in the source areas in the Semitau High (Fig. 7.26D). The decrease in coarse sandstone content towards the top of the formation and through the overlying Silat Shale indicates waning relief and a period of tectonic quiescence during the deposition of the fluvial and lagoonal to shallow marine Suwang Group. An increase in sandstone content toward the top of the Silat Shale heralds the start of another uplift phase north of the basin. Thrusting associated with this uplift is shown by the intense period of intrabasinal deformation that occurred after the deposition of the Suwang Group and resulted in the development of the tightly folded Silat Fold Belt on the northern margin of the basin. This fold belt consists of two tight east-plunging synclines with steep northern limbs, overturned in places, and gentle southern limbs. These synclines formed during southward backthrusting of the adjacent accretionary wedge.
During this period of deformation in the Late Eocene, the accretionary wedge grew northward to form the Lubok Antu and Kapuas Melange zone (Fig. 7.26F). The Lubok Antu Melange, which occupies a broad arcuate belt to the north of the Ketungau Basin, is represented by a tectonically mixed assemblage of highly cleaved and sheared Late Cretaceous to Early Eocene (Discoaster lodoensis zone, Tan, 1979) flysch deposits, pelagic sediments and mafic igneous rocks. This accretionary wedge has been extensively disrupted, folded, faulted and subjected to low grade regional metamorphism (prehnite-pumpellyite facies) during Eocene subduction of the northern oceanic floor beneath the arc system in western Kalimantan (Tan, 1982, 1986).

The new position of the Benioff zone accounts for the Late Eocene volcanism in the Semitau High and also explains the northward migration of the basal backthrust. It also resulted in uplift in southern Sarawak associated with a new rapidly subsiding forearc basin, the Ketungau Basin, developing on the continental side of the new accretionary prism, that is between the old and new outer arc ridges.

In the Melawi Basin, the relatively undeformed Melawi and Kapuas Groups unconformably overlie the Suwang Group and older rock units. The coarse fluvial deposits of the Sepauk Sandstone were derived from the southwest and gave way eastward and northwards to the marginal shallow marine portions of the Payak Formation and the brackish water to
lower floodplain and fluvial channel deposits of the Tebidah Formation (Fig. 7.26F). The main contribution from the Semitau High at this time was tuffaceous volcanic detritus that forms a distinctive facies in the Payak Formation. The Melawi Group was folded during Early Oligocene into a gentle syncline before the Sekayam and Alat Sandstones were deposited (Fig. 7.26G). The angular discordance in the north between the Kapuas Group and the Tebidah Formation decrease to zero in the core of the syncline. The Kapuas Group also extends over parts of the Semitau High indicating low topographic relief over the Boyan Melange at this time.

In contrast the younger Ketungau Basin sequence is a conformable succession. The Kantu Formation starts with basal near shore marine deposits that pass upwards into the marginal marine, estuarine and predominantly fluvial and floodplain deposits of the upper Kantu Formation. Transport directions were southwestward (Fig. 7.10) indicating derivation from uplifted area to the north of the basin (Fig. 7.26F). The overlying fluvial Tutoop Sandstone defines a syncline in the western part of the basin, but to the east both outcrop and the syncline are poorly defined. An equivalent unit (Kapuas Group) is preserved across the Semitau High indicating that activity on this structure had ceased prior the deposition of this blanket sandstone (Fig. 7.26G). Continued subsidence in the Ketungau Basin led to accumulation of further
floodplain deposits constituting the Ketungau Formation (Fig. 7.26H).

The presence of ultramafic detritus in the sediment of the Ketungau Basin and the abundance of chert and mafic grains in the Alat Sandstone, which in part overlies the Boyan Melange, indicates derivation from a Late Eocene to Oligocene structural high, the Lubok Antu Melange, to the north of Ketungau Basin. The northern margins of both the Melawi and Ketungau Basins are faulted, probably marking the position of the original basal backthrust of the relevant accretionary prism.

In conclusion sedimentation in the Melawi and Ketungau Basins was controlled by backthrusting of the Late Cretaceous to Early Tertiary accretionary wedge of Northwest Kalimantan and Sarawak. The Ketungau Basin can be correlated with the upper part of the Melawi Basin sequence and resulted from the uplift of a second accretionary wedge following the northward step of the Benioff zone. The tectonic setting of these basins was interpreted by Pieters and Supriatna (1989) as foreland basins, whereas the interpretation in Katili (1973, 1975, 1989), Hamilton (1979), Williams et al. (1988) and in this thesis is that they developed as forearc basins.
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CHAPTER 8
CONCLUSIONS

1. Integration of palynological, sedimentological and petrological data has necessitated a revision of the stratigraphic succession in the Melawi Basin and provided positive correlation between the Melawi and Ketungau Basins (Fig. 2.3). The following units are recognized in the Melawi Basin: the basal Ingar Formation overlain successively by the Suwang Group (Dangkan Sandstone and Silat Shale), the Melawi Group (Sepauk Sandstone, Payak and Tebidah Formations) and Kapuas Group (Sekayam and Alat Sandstones). The base of each group is marked by an unconformity. In the Ketungau Basin sequence, the Merakai Group shows an essentially conformable sequence and extends from the Kantu Formation up through the Tutoop Sandstone and Ketungau Formation.

2. Based on the Q-mode cluster analysis dendogram the sandstone from the lower part of Ketungau Basin (Kantu Formation) has a close relationship with the sandstone from the middle part of the Melawi Basin (Melawi Group), and the Tutoop Sandstone (Ketungau Basin) is petrographically similar to the Alat Sandstone (Melawi Basin). These units also show similar diagenetic stages. Previous nomenclature had suggested that the Tutoop and Dangkan Sandstones were equivalent.
3. The sequence of depositional environments in the Melawi and Ketungau Basins is based on the contained sedimentary structures, fossils and macerals in each formation. The Melawi Basin sequence began with the deposition of the Ingar Formation in a deep outer shelf to upper slope environment. It was followed, after a period of uplift and erosion, by the low sinuosity fluvial deposits of the Dangkan Sandstone and the lagoonal to marine trough and marginal marine deposits of the Silat Shale. After a second period of folding concentrated along the northern margin of the basin, the Payak Formation was deposited in a terrestrial to shallow marine environment in the northern part of the basin while to the south the Sepauk Sandstone was deposited as fluvial channel to nearshore marine sands. These units partly interdigitate with the marginal marine deposits of the lower Tebidah Formation which passes gradationally up into floodplain and isolated channel sand deposits. After a further tectonic episode the Sekayam Sandstone was deposited in a high energy fluvial environment in the western part of the basin while the Alat Sandstone was deposited by low sinuosity to meandering streams in the eastern part of the basin.

The Ketungau Basin sequence began with the shallow marine, nearshore and floodplain deposits of the Kantu Formation, followed by the braided to low sinuosity fluvial Tutoop Sandstone and ended with the floodplain to marginal marine deposits of the Ketungau Formation.
4. Palaeocurrent directions for the sandstone units in the Melawi and Ketungau Basins (Figs 7.1, 7.3, 7.5, 7.8 and 7.10) indicate a dominant northerly source area, with minor contributions from the Schwaner Block to the south. Petrologically determined provenance and tectonic setting of the sandstone in the Melawi and Ketungau Basins (Tables 4.5, 4.6 and 4.7) indicates that it was mainly derived from recycled orogenic material in the Semitau High, i.e. from the Boyan Melange and Selangkai Formation. Only the sandstone in the Ingar Formation was derived from a magmatic arc source (the Schwaner Mountains) to the south of the basin. The sandstone in the Ketungau Basin was entirely derived from the Lubok Antu Melange to the north of the basin.

5. Based on diagenetic features the sandstone in the Melawi and Ketungau Basins can be differentiated into three diagenetic stages. Firstly, the early shallow burial diagenetic stage is represented by Ketungau Formation, Tutoop Sandstone and Kapuas Group with burial temperatures of about 75°C, a pressure of about 37.5 MPa and a depth of burial of about 3750 m. It is characterized by weak compaction and the presence of clay grain coatings. The second diagenetic stage is represented by the Kantu Formation and Melawi Group with a depth of cover of about 6750 m, a temperature of about 80°C and a pressure about 135 MPa. It is
characterized by the presence of common secondary porosity. The third diagenetic stage has a depth of burial of about 9750 m, a temperature of 82°C and a pressure of 240 MPa. It is characterized by the presence of laumontite, quartz and calcite cement and by extensive pressure solution. The temperatures for the second and third diagenetic stages were probably higher in the central parts of the basins where depths of cover were greater.

6. The vitrinite reflectance in the Melawi Basin sequence ranges from approximately 0.51% $R_{v,\text{max}}$ in the Alat Sandstone to approximately 0.68% $R_{v,\text{max}}$ in the Ingar Formation, while in the Ketungau Basin sequence the vitrinite reflectance varies from 0.66% $R_{v,\text{max}}$ to 0.78% $R_{v,\text{max}}$. Temperatures indicated by vitrinite reflectance range from 75°C to 82°C. High reflectance values found in some beds range from 0.94% $R_{v,\text{max}}$ to 3.22% $R_{v,\text{max}}$ in the Melawi Basin and from 0.86% $R_{v,\text{max}}$ to 1.54% $R_{v,\text{max}}$ in the Ketungau Basin. The anomalous values are attributed to heating associated with the igneous Sintang Intrusive. Vitrinite is the most abundant maceral group within both the coal and shaly coal lithologies in both the Melawi (82.40±12.01%) and Ketungau (96.14±1.52%) Basin sequences. It consists dominantly of telovitrinite and detrovitrinite. The liptinite maceral group in both basins includes cutinite, liptodetrinite, resinite, sporinite and
bituminite. Inertodetrinite is the dominant inertinite maceral.

7. The best quality coal, with the greatest potential for mining in the Melawi and Ketungau Basin sequences is found within the Alat Sandstone (Alat 1 Seam). The possibility of the formation and accumulation of hydrocarbons in the Melawi Basin is highest in the anticlinorium area around the middle course of the Kayan River, where the Tebidah mudstone could act as a cap rock, the Payak Formation as a potential reservoir, and possible source rocks include the Ingar Formation, Silat Shale and the Tebidah Formation itself.

8. The tectonic history of northwestern Kalimantan (Fig. 7.26) began in the Permian but Triassic and Jurassic units are sparse. Late Cretaceous subduction produced the Boyan Melange and Permian granitic microcontinents were incorporated into this melange belt. The Semitau High was uplifted during the Paleocene or Early Eocene due to motion along a set of backthrusts. The uplifted accretionary prism was flanked to the south by the developing forearc Melawi Basin. Sedimentation and erosion events (unconformities) in the Melawi Basin were controlled by backthrusting and uplift of the Late Cretaceous to Early Tertiary accretionary wedge to the north of the basin. The northward migration of the Benioff Zone in the Late Eocene resulted in a second
accretionary wedge (the Lubok Antu Melange of northwest Kalimantan and Sarawak) and created a second forearc basin (the Ketungau Basin) between the old and new outer arc ridges. Sediment supplied from the younger outer arc ridge spread southward through the Ketungau Basin to the Melawi Basin over the earlier outer arc ridge.
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