PETROGENESIS OF THE
PENINSULAR MALAYSIAN
GRANITOID BATHOLITHS

by

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STATEMENT

This thesis is based on experimental work carried at the Research School of Earth Sciences and the Department of Geology, Australian National University between October, 1979 and June, 1983.

The x-ray fluorescence and instrumental neutron activation analyses were the result of a joint effort between B. W. Chappell and myself. Unless otherwise acknowledged, all other analytical work and the conclusions presented in this thesis are those of the author.

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ABSTRACT

A north-south trending medial suture separates the East Coast and West Coast granitoid provinces of Peninsular Malaysia. U-Pb zircon dating of selected plutons from the West Coast Province yields precise and unambiguous ages of 198 to 220 Ma (late Triassic). These ages are considerably older than K-Ar mica ages which have been largely reset but agree with or are younger than Rb-Sr whole rock apparent ages. The U-Pb zircon ages are considered to represent best estimates of the intrusive ages of the plutons. Pre-Triassic Rb-Sr whole rock apparent ages are believed to be an artifact of fractionation-assimilation processes. Reverse discordant patterns displayed by zircon populations from a number of plutons yield consistent upper intersection ages of ~1500 – 1700 Ma that plausibly date source-derived, inherited zircon components. The Precambrian ages of these inherited zircon components indicate the presence of old Precambrian crystalline basement underlying or adjacent to the West Coast Province.

U-Pb zircon dating of plutons from the East Coast Province of Peninsular Malaysia confirms the Permian to Triassic ages of the granitoids of this province indicated previously by K-Ar mica and Rb-Sr whole rock methods. Inherited zircon ages of ~800 Ma and ~1350 Ma are indicated by reverse discordant patterns of two plutons.

The late Triassic granitoids of the West Coast Province constitute an exclusively S-type plutonic province interpreted to be related to continent-collision. Petrographic and chemical characteristics of representative suites of these S-type granitoids are discussed. The chemical data are used to semi-quantitatively model crystal fractionation processes and describe aspects of late
stage fluid phase interactions. Inherited zircon, Nd and Sr isotopic results indicate that the granitoids were derived by partial melting of sedimentary rocks recycled from old Precambrian basement.

Three distinct plutonic associations are recognized in the Permian-Triassic subduction-related East Coast Province - a dominant I-type granitoid association, a less voluminous S-type association and a minor mafic association. The East Coast plutonic province is bimodal in nature - there is a SiO₂ gap between 55 to 65% SiO₂ that reflects the scarcity of intermediate rocks. Semi-quantitative petrochemical modelling results are used to constrain a discussion of I- and S-type fractional crystallization processes. The S-type granitoids of the East Coast Province display chemical and isotopic (Nd and Sr) signatures that are compatible with derivation from sedimentary rocks recycled from Precambrian crystalline basement. Multi-stage processes are invoked to explain the genesis of the I-type granitoids. Partial melting of juvenile (1st stage mantle derivation) underplated crust and old lower crust of mafic-intermediate compositions, supplemented to various degrees by crystal fractionation, hybridization and assimilation provides the framework to interpret the complex chemical and isotopic signatures of the I-type granitoids. The mantle-derived mafic (mainly gabbroic) rocks display the broad characteristics of orogenic arc tholeiites.

Results of an integrated U-Pb zircon and Nd isotopic study have allowed first order age estimates of the continental blocks underlying the East Coast and West Coast Provinces of Peninsular Malaysia to be made. The results indicate 'crust formation' ages of ~ 1500 – 1700 Ma and ~ 1100 – 1300 Ma for the West Coast and East Coast crustal segments respectively. It is observed that crustal segments
underlying many major Phanerozoic continental margin volcano-plutonic provinces consistently yield mid-Proterozoic Nd model ages that plausibly represent 'crust formation' ages. The interpretation that many continent margins display mid-Proterozoic 'crust formation' ages, if correct, indicate that many continental blocks reached very much their present day extent in the mid-Proterozoic. Phanerozoic arc magmatism does not appear to be a major mechanism for crustal growth. The ubiquity of mid-Proterozoic 'crust formation' ages is suggestive of an important global mid-Proterozoic crustal addition episode(s).
# Table of Contents

**Acknowledgements**

**Abstract**

**Table of Contents**

**Chapter 1 - Introduction**

**Chapter 2 - Geology of Peninsular Malaysia**

**Chapter 3 - A U-Pb Zircon Geochronological Study of the Peninsular Malaysian Granitoids**

- Introduction
- Experimental techniques
- Previous work
- Granitoids of the West Coast Province
- Granitoids of the East Coast Province
- Aspects of zircon inheritance and Zr saturation
- Conclusions

**Chapter 4 - Granitoid Batholiths of the West Coast Province of Peninsular Malaysia**

- Introduction
- Experimental techniques
- Geological and tectonic setting
- Batholith constitution and suite recognition
- General major element chemistry
- Mineral chemistry
- Major and trace element chemistry
- High-SiO₂ granites
- Coeval volcanism
- Isotopic characteristics
- Summary
CHAPTER 5 - GRANITOID BATHOLITHS OF THE EAST COAST PROVINCE OF PENINSULAR MALAYSIA

Introduction 121
Tectonic and geologic setting 122
Batholith distribution and age patterns 123
Experimental techniques 124
General major element chemistry 125
Mineral chemistry 126
Major and trace element chemistry 134
Rocks of the mafic batholith association 150
Isotopic relationships 159
Synthesis 163

CHAPTER 6 - GENESIS OF GRANITOID BATHOLITHS OF PENINSULAR MALAYSIA AND IMPLICATIONS FOR MODELS OF CRUSTAL GROWTH: EVIDENCE FROM A Nd - Sr ISOTOPIC AND U-Pb ZIRCON STUDY

Introduction 170
Experimental techniques 179

East and West Coast granitoid provinces 180
West Coast Province 182
East Coast Province 187
Nd isotopic characteristics of Phanerozoic margin magmatic belts 192
Conclusion 195

CHAPTER 7 - SUMMARY AND SYNTHESIS 198

APPENDIX 1 - The U-Pb zircon method 215
APPENDIX 2 - The Sm-Nd method 222
APPENDIX 3 - The Rb-Sr method 229
APPENDIX 4 - X-ray fluorescence spectrometry 232
APPENDIX 5 - Instrumental neutron activation analysis 234
APPENDIX 6 - Table of chemical analyses 236
APPENDIX 7 - Sample locations and descriptions 251
REFERENCES 268
INTRODUCTION

The broad aim of this thesis is to provide an integrated mineralogical, chemical and isotopic characterization of the granitoid batholiths of Peninsular Malaysia. In this introductory chapter, a number of important developments in granitoid petrology are reviewed in order to provide a broad framework for an appreciation of aspects of the petrogenesis of the Peninsular Malaysian batholiths discussed in subsequent chapters.

Phanerozoic granitoids can be found in a variety of tectonic environments, ranging from subduction-associated island arc and Andean-type settings to rift-related regions. Granitoids belonging to the orogenic plutonic association are located in regions where they are related to ongoing plate convergence and broadly synchronous folding and regional metamorphism. A second group considered to belong to an anorogenic plutonic association may develop very late in a magmatic cycle (e.g. A-type suites of Bega Batholith (Collins et al., 1982)) or are found in tensional regimes associated with minor mafic magmatism (Nigerian Mesozoic complexes, van Breemen et al. (1975)), or associated with voluminous mafic intrusive–volcanic complexes (e.g. Tertiary Hebridean Province of Scotland, Moorbath and Thompson (1980)). This thesis will only concern itself with granitoids of the orogenic plutonic association. Occurrences of such granitoids range from small discrete plutons in young island arc terrains to large complex composite batholiths along continent margins.
Granitoid batholiths are a distinctive feature of Phanerozoic fold belt systems and sometimes exceed fifty percent of the exposed area of a fold belt. Their extremely variable mineralogical and chemical characteristics, differences in depths of emplacement and thermal environments, variable grade of associated metamorphic rocks and poorly understood relationships with coeval volcanism make simple generalizations concerning the origins and evolution of the granitoid batholiths of Phanerozoic fold belts extremely difficult. Workers like Pitcher (1979) have both emphasized the uniqueness of every fold belt and stressed the need to appreciate and document both the differences and similarities of the granitoid batholiths found in various terrains.

**Granitoids and continental crust**

One common denominator of the various granitoid types is their association with continental crust, be it old Precambrian crystalline basement or young island arc crust. Voluminous granitoid batholiths are only found along the margins of large continental crustal blocks. The intimate relationship between granitoids and continental crust may result from two contrasting modes of origin - (1) granitoids are primarily crustal remelts or (2) the deep continental crust represents an essential staging area where mantle-derived mafic magmas can stagnate and differentiate to granitoid compositions. The first mode of origin implies that granitoids do not represent new additions to the continental crust but that plutonism only affects a vertical chemical redistribution within the crust. Partial melting extracts a melt phase enriched in incompatible elements and leaves behind a depleted mafic residue. The second model represents the case where granitoids are taken to be juvenile additions to the crust and argues
that emplacement of granitoid batholiths is the primary mechanism by which continental crust grows with time. These simple models invariably obscure important aspects of granitoid genesis. In fact, the complex chemical and isotopic evidence emerging suggest that both models may need to be invoked for a single plutonic-volcanic terrain (e.g. Hawkesworth et al., 1982). It is increasingly obvious that detailed case histories of individual plutonic provinces are required before any simple generalization can be made concerning crustal vs mantle inputs. This thesis is a small step in this direction.

The plate tectonic connection

In this thesis, the basic tenets of plate tectonic theory as summarized by Le Pichon (1968), Isacks et al., (1968), Dewey and Bird (1970) and by many others are assumed. Although details still need to be worked out, the evidence in favour of convergent plate interactions as a first order mechanism that leads to the generation of granitoid magmas of the orogenic association is now well established. The need to relate Phanerozoic granitoid magmatism to more detailed plate tectonic environments is, I think, an important and worthwhile challenge facing granitoid petrologists. Major hurdles confront attempts to unravel tectonic environments older than the Tertiary and almost inevitably, arguments based on structural, metamorphic and sedimentological grounds need to be utilised in an integrated fashion. Pitcher (1979), for example, described three broad structural-tectonic environments (Alpinotype, Andinotype and Hercynotype based on their various type settings) in which granitoids are found and he expressed the need for geochemical investigations to be pursued in terms of tectonic models to produce more realistic hypotheses of granitoid genesis.
Five broad granitoid environment associations are generalized in Fig. 1 which is partly based on the model of Pitcher (1979). Combinations of different associations in a single complex terrain may be possible. In order of increasing continental crustal input, the five associations may be listed as:

1. granitoid plutons in primitive island arcs like New Britain (Whalen, 1980),

2. plutons intensive into the accretionary wedge in fore-arc settings adjacent to continent margins e.g. southwest Alaska (Hill et al., 1981),

3. continental margin cordilleran granitoid batholiths like the Peruvian Coastal Batholith (Pitcher, 1978; Atherton et al., 1979; McCourt, 1981),

4. intracontinental batholiths and plutons like the Inner Cordilleran Belt of peraluminous plutons described by Miller and Bradfish (1980) in western U.S., and

5. continental collision plutons as exemplified by the Tertiary Himalayan leucogranites (Le Fort, 1981; Dietrich and Gansser, 1981).

It is stressed that the groupings above do not represent an attempt at 'pigeon-holing' granitoids. They represent what are regarded as rather unambiguous examples of Tertiary granitoid magmatism at different tectonic settings for which data are available for comparative granitoid studies. Studies documenting the similarities and differences in the characteristics of granitoids at
Fig. 1. Highly schematic environment settings for granitoids of the orogenic association. Subduction relationships are idealised - although spatial and temporal proximity to subduction are well established, exact causes and relationships between plutonism and subduction are complex. Complications also arise due to tectonic superimposition of different plutonic belts and trench migration. Superimposition of plutonic belts may explain the presence of only S-type granitoids in some accretionary wedge settings and the entire gabbro to granite range in others.

Plutonism associated with hypothetical situations involving arc-arc or arc continent collisions may be considered as arc or continent margin environments depending on whether old crust is involved or not. The suggested scheme represents only a first order approximation to distinguish granitoid environments. Deviations from these broad environments are possible.
Island arc
- e.g. New Britain
- I-type plutons
- gabbro to granite, diorite-tonalite main rock types
- coeval calcalkaline volcanics
- batholiths absent

Accretionary wedge
- e.g. southwest Alaska
- I-, S-type plutons
- gabbro to granite, tonalites common
- coeval calcalkaline volcanics
- batholiths absent

Continent margin plutonic arc
- e.g. Coastal Batholith of Peru
- I-, S-type plutons
- gabbro to granite, tonalite-granodiorite main types
- coeval calcalkaline volcanics
- batholiths absent

Intracontinental plutonic belt
- e.g. Idaho Batholith, Western U.S.A.
- S-type plutons
- gabbro to granite, granodiorite to granite, rare diorite
- coeval calcalkaline volcanics
- voluminous linear batholiths

Continent collision plutonic belt
- e.g. Himalayan
- Tertiary leucogranites
- S-type plutons
- granodiorite to granite
- minor silicic volcanics
- silicic volcanics absent or very minor
- batholiths present or may be present

Diagram showing continental crust and upper mantle.
similar settings would facilitate a better understanding of the igneous processes that lead to 'chemical variations on a common theme'.

**Composite nature of granitoid batholiths**

The recognition that large complex batholiths can be systematically divided into a small number of suites (or sequences, series, superunits) (Bateman and Dodge, 1970; Cobbing and Pitcher, 1972; White et al., 1977) with specific mineralogical, textural and chemical signatures has greatly aided the deciphering of the constitution of such batholiths. The principles by which suites can be recognized are outlined by Pitcher (1978). A single intrusive suite consists of a variable number of consanguineous cooling units or plutons. Suites may also be more loosely combined to form groupings called supersuites. Presnall and Bateman (1973) presented arguments to show that suites may be interpreted as representing high-level differentiation of a parental magma formed during a single fusion episode.

It will be argued in this thesis that the concept of suite division can be applied to unravel the complex structure of the Peninsular Malaysian batholiths. Independent confirmation of this seems to be indicated by the preliminary results of a new field mapping study by geologists of the Institute of Geological Sciences of London and the Geological Survey of Malaysia published in a very short note by Cobbing et al. (1982).
Granitoid magmas

There is presently little support for mechanisms based largely on in-situ metasomatic processes to produce the large granitoid batholiths found along continent margins. The present consensus is that granitoid magmas are involved (e.g. chapter 12 of Carmichael et al., 1974; White and Chappell, 1977; Barker, 1981; Hildreth, 1981). Problems regarded as of crucial importance to the understanding of the origins of granitoid magmas are:

1. the nature of the sources of the various granitoid magma types,
2. the nature of magma generation processes, and mechanisms of ascent and emplacement of granitoid plutons,
3. the nature of differentiation processes yielding the large chemical variations observed in granitoid suites.

Problems 1 and 3 are amenable to study using integrated chemical and isotopic data. Crustal, mantle and hybrid origins of granitoid magmas have been proposed to explain the chemical and isotopic characteristics of various granitoid types. Hybrid origins in particular raise the difficult problem of estimating the proportions of crustal and mantle components involved. Studies attempting to differentiate between crustal versus mantle sources of granitoid magmas led to the gradual recognition of two main orogenic granitoid associations (Pitcher, 1979) - one, a compositionally expanded series ranging from gabbro to granite and another, a compositionally restricted series ranging from granodiorite to granite.
Chappell and White (1974) and later, Hine et al. (1978) and McCulloch and Chappell (1982) showed that the early Palaeozoic batholiths of southeastern Australia could be divided into two contrasting genetic associations. Marked chemical and isotopic differences shown by the two granitoid associations led these workers to discuss their chemical and isotopic data in terms of a very explicit binary division of granitoid types - an S-type association derived from partial melting of metapelitic source rocks and an I-type association derived from igneous source rocks that may or may not be of first-stage mantle origin.

S-type granitoids described by these workers typically contain the ferromagnesian assemblage high-Al biotite + muscovite + cordierite + garnet. Sillimanite may be a common accessory and ilmenite is the typical opaque oxide. The rocks are invariably strongly peraluminous with mol. $\frac{Al_2O_3}{(Na_2O + K_2O + CaO)} > 1.1$, show a restricted $SiO_2$ range, have high $K_2O$ and low $Na_2O$ contents and possess low $Fe_2O_3/FeO$ ratios. Sr initial ratios are typically greater than 0.710, initial $\varepsilon_{Nd}$ values are characteristically negative (-6 to -10) and $\delta^{18}O$ exceed + 9.5. I-type granitoids, in contrast, contain low-Al biotite + hornblende + sphene. Magnetite is the typical opaque oxide. I-types are generally metaluminous with mol. $\frac{Al_2O_3}{(Na_2O + K_2O + CaO)} < 1.1$. They display a wide range in $SiO_2$ contents, have low $K_2O$ and high $Na_2O$ contents and high $Fe_2O_3/FeO$ ratios. Sr initial ratios are less than 0.710, initial $\varepsilon_{Nd}$ values range from + 1 to -9 and $\delta^{18}O$ values range from + 8 to + 9.5.
Australian workers were further able to distinguish an 'I-S line' in southeastern Australia (White et al., 1976) that separates a belt of predominantly S-type granitoids from a solely I-type granitoid province. The 'I-S line' probably represents a fundamental change in deep crustal composition and seems to mark the most easterly margin of old underlying Precambrian crust in southeastern Australia. Other workers have also documented the importance of source control on granitoid composition. The 'quartz-diorite line' (Moore, 1959) in western U.S.A. probably has similar significance to the 'I-S line'. Kistler and Peterman (1973) observed that regional compositional changes across the Sierra Nevada Batholith of western U.S.A. are independent of age of the granitoids and probably reflect the nature of the underlying source rocks. Silver et al. (1979) concluded that high-level fractionation cannot explain isotopic patterns like the systematic correlation between Sr initial ratios and $\delta^{18}O$ across the Peninsular Ranges Batholith of California. They attributed the patterns to systematic differences in chemical and isotopic characteristics in the deep source regions. These and other studies show the potential of using granitoids as images of underlying crust.

As more and more granitoid terrains were examined in terms of an I-, S- type granitoid model, it became clear that although the basic I-S division could be widely recognized (Beckinsale, 1979; Takahashi et al. 1980; Dietrich and Gansser, 1981; Shaw and Flood, 1981), criteria for distinguishing the two granitoid types in southeastern Australia could not be literally extended to all terrains. This primarily reflects the unique characteristics of the source regions of different terrains and the unique interactions between crust and mantle in each case. In addition, Shaw and Flood (1981) recognized suites with characteristics intermediate between I- and S- type
granitoids in the New England Batholith of eastern Australia. Such intermediate suites may be derived by melting of mixed source rocks containing peraluminous and metaluminous lithologies or from immature volcanogenic sediments.

Despite such limitations, the I-S division has many advantages over other classifications for the Peninsular Malaysian context. A calcalkaline/alkali-calcic/calcic description is ambiguous because of the scarcity of samples in the compositional fields where the alkali-lime index could be defined in the manner Peacock (1931) envisaged i.e. intersection of total alkali and CaO trends plotted against SiO₂. The index strictly refers only to a single differentiation suite and is incapable of distinguishing sedimentary-derived peraluminous granites from those formed by fractionation from metaluminous parents. The latter criticism also applies to the use of simple AFM triangular plots (Fig.2) and the K₂O vs. SiO₂ classification scheme for chemically comparable volcanics (Fig.3). Streckeisen's (1973) granitoid nomenclature (Fig.4) is useful and is adopted in this thesis. It however lacks important genetic information and is combined with an I-S division to describe the granitoids of Peninsular Malaysia. This was considered to represent the best approach to adopt in view of the rather fluid state of granitoid classification schemes at present. This approach combined with tectonic environment groupings discussed earlier departs from classifications and descriptions based on the nature of the metamorphic environment, the depth of emplacement and relationships to folding phases.
Fig. 2 Compositional trends of magma suites containing mafic to felsic members. Solid line separates tholeiitic (Th) from calcalkaline (CA) suites. Dashed lines connect Kuno's (1968) pigeonitic (PRS) and hypersthenic (HRS) rock series. Diagram after Gill (1981).
Fig. 3 % K$_2$O vs. % SiO$_2$ plot (after Ewart, 1979) used to classify volcanic suites containing andesites, dacites and rhyolites.
Fig. 4 Modified Streckeisen's (1973) modal classification of granitoids based on various proportions of quartz, alkali feldspar and plagioclase. The useful term adamellite has been retained. 1, granite; 2, adamellite; 3, granodiorite; 4, tonalite; 5, quartz diorite; 6, quartz monzodiorite; 7, quartz monzorite.
Magma modifying processes

Magmatic differentiation may yield derivative compositions that are drastically different from parental magma compositions. It is of critical importance to be able to identify and isolate the effects of differentiation processes that obscure the nature of parental magmas and the sources of such magmas. Differentiation processes may in addition result in anomalous enrichments of rare metals of economic importance.

The most well-documented magma modifying process is that of crystal-liquid fractionation (Bowen, 1928) whereby liquidus crystalline assemblages are prevented from reaction with an evolving magmatic liquid by processes like filter-pressing, localization of crystallization (e.g. sidewall crystallization) and displacement of residual magmatic liquids inwards or upwards within a fractionating magma chamber. Generalized quantitative treatments of fractional crystallization behaviour are given in McCarthy and Hasty (1976) and Hanson (1978).

Restite unmixing (White and Chappell, 1977) is another possible mode of differentiation capable of generating a range of compositions. The mechanism involves the variable separation of a partial melt from the refractory residue in ascending magma diapirs.

Magma mixing, as distinct from restite unmixing, involves physical mixing of initially unrelated magma batches. It may occur at various depths in the crust. A number of workers including Eichelberger (1978) have presented evidence for variable amounts of mingling between silicic and basaltic liquids to yield a large range of intermediate rocks.
Assimilation and combined fractionation-assimilation (Bowen, 1928; Carmichael et al., 1974; DePaolo, 1981a) describe processes by which components of country rocks may be incorporated into a rising, differentiating magma by partial melting of adjacent wall rocks, by digestion and reequilibration with accidental xenoliths or by partial chemical exchange with wall rocks and xenoliths. Assimilation probably occurs largely in synchronicity with fractionation at depth, the latent heat of crystallization being a source of heat that counts heat losses during assimilation.

Liquid state differentiation is a general term that is used here to describe processes affecting compositional changes in the liquid state. Included in this category are processes like compositional convection whereby liquids of differing densities in a magma chamber are gradually separated by buoyancy forces; volatile complexing whereby transport and enrichment of rare metals are enhanced by formation of soluble complexes; thermogravitational diffusion (Hildreth, 1981) whereby compositional gradients arise between boundary layers sustained by thermal convection; and changes in melt structure with increasing differentiation that cause anomalous trace element partitioning behaviour in anhydrous, highly polymerized melts (Mahood and Hildreth, 1982). A more detailed discussion of these processes pertinent to the genesis of high-SiO₂ liquids is found in Hildreth (1981).

Fluid phase separation during the second boiling reaction:

\[ H_2O\text{-saturated melt} = \text{crystals} + \text{volatile-rich fluid phase} \]

is an important end-stage differentiation process that may be the primary mechanism for generation of ore-forming fluids associated with granitoid magmas. Physico-chemical factors involved in the generation
of a fluid phase are discussed by Jahns and Burnham (1969), Burnham (1979) and Burnham and Ohmoto (1980).

More than one of the possible processes listed above may operate at any moment in the evolution history of a granitoid body. One aim of this thesis is to identify and discuss the effects of various differentiation mechanisms affecting the Peninsular Malaysian granitoids.

The rheologic behaviour of granitoid magmas assumed in this study is based on the studies and reviews by Van der Molen and Paterson (1979), Pitcher (1979), Marsh (1981) and others cited by these workers. Granitoid magmas are believed to display a range of rheologic properties throughout their magmatic histories - ascending crystal-rich magma diapirs may contain as little as 30 to 50% melt component and behave as pseudo-plastic Bingham bodies but differentiation processes may yield largely liquid, volatile-rich and less viscous end-stage differentiates. Gravity settling is assumed to be an unrealistic mechanism; thus, 'hidden cumulate' fractionation models are regarded as unlikely. Marginal accretion of solids and displacement of liquids inwards and roofwards may be the main mechanism by which magmas cooled and solidified, with possibly late surges and injection of residual liquids disrupting simple cooling sequences (e.g. Bateman and Chappell, 1979).

**Constraints from experimental petrology**

An important aspect of igneous petrology is the identification of the nature and sources of primary magmas. Although experimental petrological studies cannot identify primary magmas per se, phase equilibria studies at low-moderate pressures allow insights into
probable equilibria during crystallization and differentiation processes that alter the compositions of primary magmas whereas phase studies at high pressures allow insights into deep-seated, melt generation equilibria. Experimental work designed specifically to answer questions regarding the origins and crystallization behaviour of granitoid magmas include the works of Tuttle and Bowen (1958), Wyllie et al. (1976), Wyllie (1977), Clemens and Wall (1981) and many others.

Wyllie et al. (1976), for example, used their data to constrain "possible and impossible sources" of granitoid magmas, their water contents and crystallization sequences. They argued that the widespread cordilleran I-type plutons of tonalitic composition cannot be generated as primary magmas from mantle peridotite or by ultrametamorphism of the lower crust unless significantly higher temperatures than those of regional metamorphism are attained. These workers suggested that it was more probable that tonalitic magmas represent profoundly modified and fractionated magmas formed by complex interactions between crustal- and mantle-derived magmas. Gill (1981) came to rather similar conclusions concerning the origins of andesites. Clemens and Wall (1981) described the phase behaviour for an S-type granitoid and used their data to discuss stabilities of mineralogical assemblages during partial melting and high-level crystallization of S-type magmas, estimated water contents of S-type magmas and explained how differing chemical features of S-type suites could be generated by different mica breakdown reactions at varying pressures and temperatures.
Geochemical and isotopic models developed in chapters four and five of this thesis to explain the petrogenesis of the Peninsular Malaysian granitoids are constrained, where not explicitly stated, by the broad conclusions of these experimental studies with respect to probable fractionating assemblages, source mineralogy and range of P-T conditions. The results of such experimental petrological studies have allowed the development of simpler models and the exclusion of otherwise cumbersome and involved discussion of alternative models.

Impact of the application of isotopes

The use of isotopic tracers to place constraints on granitoid sources has been one of the most important recent advances in granitoid petrology. In this respect, the use of Sr isotopes has been notably widespread and successful. In recent years, the application of multiple isotope tracing has resulted in great strides towards the goal of deciphering the multiple origins of granitoid magmas. Hart and Allègre (1980) summarized Nd, Sr and Pb isotopic data for the French Hercynian granitoids, DePaolo (1981b) described relationships between Nd, Sr, O and Pb isotopes for the Sierra Nevada Batholith of western U.S.A. and McCulloch and Chappell (1982) summarized the Nd, Sr and O data for the Berridale-Kosciusko Batholiths of southeastern Australia. In addition to the use of these isotopes to estimate crustal and mantle components in granitoid magmas, the use of combined U-Pb zircon and Nd studies has allowed geochronological as well as geochemical information pertaining to the nature of granitoid source regions to be obtained.

Just as importantly, results of stable isotope studies using O and H (Taylor, 1977) have provided insights into granitoid sources as well as interactions between magmas, aqueous magmatic-derived fluids
and meteoric hydrothermal fluids during the late-stage histories of granitoid plutons.

Harrison and McDougall (1980) discussed how a plot of a variety of radiometric ages versus closure temperatures could be used to great advantage to describe the cooling histories of granitoid plutons. Mattinson (1978) used the isotopic characteristics of plutons as well as their cooling histories to match detached blocks in western U.S.A.

Recent developments in relation to the K-Ca method (Marshall and DePaolo, 1982), Lu-Hf (Patchett, 1983), Xe (Staudacher and Allègre, 1982) and Re-Os (Allègre and Luck, 1980) systems may shed even more light on granitoid petrogenesis.

Some aspects of this thesis

The developments in granitoid petrology summarized above provide the basic interpretive framework on which this thesis is based. This study is an attempt at answering or partly answering the main issues and possibilities raised in the various sections. These are the problems of:

(a) providing a chemical and isotopic characterization of the granitoid batholiths of Peninsular Malaysia,
(b) placing constraints on the origins of the granitoid magmas,
(c) examining the characteristics of the granitoids in relation to their tectonic settings,
(d) describing the nature and extent of the various differentiating processes affecting the granitoids,
(e) using the granitoids as images of underlying continental crust.
Throughout this thesis I have tried to adopt a broad, integrated view that combines field, mineralogical, chemical and isotopic data together. In view of the reconnaissance nature of the study, I have tried as much as possible to avoid details of local significance although in certain instances, this may have been done at the expense of what some workers might regard as of potential petrogenetic significance.

**Thesis organization**

To a large extent, chapters 3-6 of this thesis have been prepared with publication as individual papers in mind and they are essentially self-contained. As such, a number of introductory descriptions and discussion of techniques have been repeated in certain sections.

Chapter 2 reviews the tectonics and geological setting of Peninsular Malaysia in the context of the evolution of the Southeast Asian region. It also documents the extent of previous knowledge pertaining to the nature and significance of the Peninsular Malaysian granitoids.

Chapter 3 is primarily a geochronological study. It discusses some of the problems involved in previous interpretations of the geochronology of the granitoids and presents the results of U-Pb zircon work for a number of critical suites. The main significance of chapter 3 is the unambiguous geochronological information obtained from the U-Pb zircon work that has (1) allowed a clearer understanding of previous Rb-Sr and K-Ar ages to emerge, and (2) provided the basic age framework on which the subsequent isotopic tracer studies are based.
Chapters 4 and 5 discuss the petrogenesis of the West Coast and East Coast Province batholiths of Peninsular Malaysia respectively in the light of new mineralogical, chemical and isotopic data.

Chapter 6 examines the significance of the isotopic results in terms of crustal imaging studies. Nd and Sr initial ratios and U-Pb zircon inheritance features are used to estimate the nature and ages of the underlying crustal blocks of Peninsular Malaysia. The results are discussed in a general context by comparison with other batholith terrains. A number of basic implications for the evolution of continental crust are then examined.

Chapters 3-6 were written assuming that the reader has prior familiarity with the behaviour of the U-Pb zircon, Sm-Nd and Rb-Sr systems, isotopic nomenclature and specialized notations. Readers not familiar with such aspects should first read the brief outlines written in appendix 1.1-1.3.

Chapter 7 offers a brief summary and discussion of the main conclusions of this thesis. Analytical techniques, location and petrographic information on analysed samples, tables listing the complete chemical data and an explanation of the isotope systems used may be found in the appendices.
Tectonics of Southeast Asia

Mainland Southeast Asia (Fig. 1) is used here to describe a region believed to be underlain by Palaeozoic or older continental crust that extends latitudinally from Sumatra into northern Burma and meridionally from eastern Burma to Indochina. Numerous accounts of the geology and tectonics of this region has been published in recent years. These include works by Stauffer (1974), Hutchison (1975, 1977, 1978), Workman (1977), Mitchell (1977, 1981) and Hamilton (1979). The review discussed below is based on these published works.

The pre-Cenozoic core of mainland Southeast Asia is bounded to the south and west by the active subduction system exposed on the surface along the Anadman-Sumatra-Java trench. The south China craton to the north is separated from Southeast Asia along the Red River suture which marks the Triassic collision of Indochina with China. Subduction complexes of mainly Jurassic - Cretaceous age indicated by ophiolite assemblages and plutonic arc rocks mark the eastern margins of the pre-Cenozoic core.

A key pre-Cenozoic geologic feature of mainland Southeast Asia is the discontinuous belt of Palaeozoic ophiolite-melange rocks ('ophiolite line' in Fig.1) extending medially through Thailand and Peninsular Malaysia. This belt has been termed the Bentong-Raub (Peninsular Malaysia) and Uttaradit - Luang Prabang (Thailand) ophiolite line (Hutchison, 1975) and Chieng Rai - medial Malaya zone (Mitchell, 1981). This ophiolite 'line' almost certainly represents subduction melange and marks the boundary between two geologically
Fig. 1. Generalized relationships of the pre-Cenozoic core of Southeast Asia to selected geotectonic features of the region. Data sources include Stauffer (1974), Workman (1977), Hutchison and Taylor (1978) and Hamilton (1979).
contrasting provinces. To the west, stretching from Sumatra to northern Burma extends a belt of Cambrian to Triassic sedimentary formations of miogeoclinal affinities i.e. non-volcanic continental platform and basin sedimentary facies, deposited along the margins of an older continental crustal basement. This stratigraphic argument indicates an adjacent Precambrian cratonic source to the present-day west of Peninsular Malaysia from Cambrian to Triassic times. East of the ophiolite - melange 'line' is a zone lacking in Lower Palaeozoic sediments and dominated by Upper Palaeozoic - Triassic intermediate to silicic plutons and volcanics, volcanoclastics and shallow water sediments. Northeast of this zone lies the Indochinese craton whose Precambrian Kontum massif represents crystalline basement.

The deduction of a former western parent craton, if correct, indicates that the pre-Cenozoic core of Southeast Asia was formed by the collision of two continental blocks - a large, unidentified western craton and an eastern block comprising Indochina, eastern Peninsular Malaysia, western Borneo and now submerged continental areas lying between these regions.

There still remain some differences between various tectonic interpretations resulting from uncertainties in the understanding of the regional geology, especially the intrusive ages of the granitoid batholiths and their tectonic settings. The plate tectonic reconstruction adopted in this study follows that of Mitchell (1977) primarily because the new geochronological and geochemical results discussed in chapters three to six of this thesis are most compatible with his model. Eastward subduction (Fig.2A) along a trench system represented by the Bentong-Raub and Uttaradit - Luang Prabang 'ophiolite line' from late Palaeozoic to Triassic is indicated by the
Fig. 2. Schematic Permian to Cretaceous evolution of mainland Southeast Asia (after Mitchell, 1977). Tectonic sections are taken along an east-west (present day coordinates) transect across the Thai-Malay Peninsula.
relationships of the ophiolite-melange belt and the intermediate to silicic volcano-plutonic are rocks found in eastern Peninsular Malaysia and eastern Thailand (Workman, 1977). Continental collision (Fig.2B) occurred during the late Triassic after the ocean or marginal basin separating the two continental masses closed by subduction of oceanic lithosphere. The age of the collision event was late Triassic because the Jurassic is characterized by the formation of widespread, undeformed flat-lying continental red bed and evaporite sequences in the Korat Plateau of Thailand and in Peninsular Malaysia. Plutonic and volcanic rocks with arc affinities of Jurassic age are absent. These observations suggest uplift and cessation of both subduction tectonics and regional deformation during the Jurassic. The Jurassic was also the time when the western parent craton was detached from the Southeast Asian region because from the Cretaceous onwards, subduction melange and an igneous arc of Cretaceous-Tertiary age developed along the western margins of the pre-Cenozoic core (Fig.2C). Seafloor data for the ocean segment west of Sumatra (summarized by Hamilton, 1979) show that oceanic crust of Cretaceous-Tertiary (50 to 90 Ma) age and possibly older has subducted beneath the Andaman-Sumatra trench.

The mechanism involved in the detachment of the western craton during the Jurassic is uncertain. Presumably, if rifting was involved, the oceanic crust formed would have been subsequently consumed by subduction beneath Eurasia during the northward advance of India during the Tertiary as indicated by seafloor spreading ridge patterns in the Indian Ocean (see Hamilton, 1979, p. 61). An alternative is detachment of the parent western craton along megashears like the Tertiary examples in western U.S.A. Perhaps the major post-Triassic strike-slip faults documented by Stauffer (1968)
and Shu (1969) represent secondary splays of such megashear systems.

Peninsula geology

A descriptive account of the geology of Peninsular Malaysia based on a compilation of the works of geologists from the University of Malaya, Geological Survey of Malaysia and Universiti Kebangsaan, Kuala Lumpur is presented in a 'Geology of the Malay Peninsula' volume edited by Gobbett and Hutchison (1973). Less involved outlines of the geology of Peninsular Malaysia can be obtained from works by Bignell and Snelling (1977) and Chung and Yin (Annual Report of the Geological Survey of Malaysia, 1968-1979). To avoid excessive duplication, the geological summary written in the next section is a brief consideration only of the aspects of these works that are relevant to a discussion of the geological settings of the granitoid batholiths.

Following Stauffer (1974) and Hamilton (1979), Peninsular Malaysia is here divided into an eastern block and a western block separated by a medial suture. Stauffer (1974) called these blocks the East and West Malaya Blocks and Hamilton (1979) used Western and Eastern Malay Peninsula Blocks.

In the western block, the oldest rocks are Cambrian sandstone-shale sequences of fluvial-deltaic origin. These are best developed on islands off the northwest coast of the peninsula and have been called the Machinchang Formation. Conformably overlying this formation is a thick shelf limestone unit of Ordovician to Devonian age known as the Setul Formation. The Machinchang and Setul Formations correspond to a continental shelf facies assemblage deposited on a stable Atlantic-type continent margin. Lying to the east of this shelf, sequences are deeper water basin facies
(shales-siltstones-cherts) of Silurian - Devonian age. Thick flysch-type sequences make up the Carboniferous Singa Formation. These pass into Permian limestones in the northern portion of the peninsula. In the central portion of the western block, deposition of limestone was essentially continuous from Silurian to Permian. The exposed Lower Palaeozoic rocks of the southern portion of the western block have been regionally metamorphosed and comprise quartz-mica schists and graphitic schists overlain by marbles of Silurian age. Clastic sediments, possibly correlative with the Carboniferous Singa Formation, unconformably overlie the Lower Palaeozoic sequences. The overall stratigraphy of the western block probably represents the superimposition of a middle Devonian - Permian foreland basin fill over a Cambrian - Devonian continental basin-shelf association (Hamilton, 1979). Intrusive into these Palaeozoic formations are large masses of granitoid batholiths yielding for the most part, late Triassic Rb-Sr isochron ages (Bignell and Snelling, 1977). Restricted occurrences of post-Triassic red beds and conglomerates suggest late Triassic uplift of the peninsula.

The narrow belt of rocks interpreted as an ophiolite-melange association separating the eastern and western blocks comprises sheared serpentinite, amphibole schist, epidote schist and blocks of chert intimately associated with highly distorted argillites, mica schists and phyllites.

East of the ophiolite-melange rocks, the stratigraphy is dominated by Carboniferous to Triassic sediments with abundant interbedded intermediate-silicic volcanics. Steeply dipping sandstone-siltstone-shale sequences and small lenses of limestone are widespread. Argillaceous facies of these sequences contain fauna,
fossil burrows, ripple marks and plant debris - all indicative of shallow water deposition. No sediments older than Carboniferous are known, but stratigraphic relationships are generally obscure. Large linear batholiths consisting predominantly of I-type plutons are exposed along the eastern margins of the eastern block. The Carboniferous - Triassic stratigraphy and igneous association can be broadly interpreted to represent a tectonic assemblage developed at a continent-margin/island arc subduction setting. Problematic occurrences of small metamorphic complexes are distributed immediately to the east of the ophiolite-melange belt. The ages of these rocks are not known and their relationships with adjacent low-grade metasediments are uncertain. It has been speculated that they might represent uplifted basement rocks. Unconformably overlying the Palaeozoic rocks is a group of fluvial-lacustrine sediments of post-orogenic origin. These scattered molasse sequences are typically flat-lying or very gently folded and are characterized by thick basal conglomerates.

Granitoid plutonism: Previous work

Results of field mapping, chemical and radiometric data up to 1973 were reviewed by Hutchison (1973a). Modal and normative plots (Fig.3) were used to show that although the granitoids range in composition from diorite to granite, most are of adamellite composition and plot near the theoretical 'granite minima' for $H_2O$-saturated liquids between 1 to 3 kb in the $Ab$ - $Qtz$ - $An$ - $Or$ tetrahedron. The early radiometric data suggested ages of emplacement for the granitoids of Carboniferous to Triassic, but small post-orogenic plutons of late Cretaceous age were also indicated.
Fig. 3. Frequency distribution of normative quartz, albite, orthoclase, and plagioclase for 119 analyzed granites and related rocks from the Malay Peninsula containing 80% or more of normative Ab + Or + Q. Contours more than 1, 2, 5, 10, and 15% based on a 0.25% counting area (after Hutchison, 1973a).
Bignell and Snelling (1977), based on the results of their Rb-Sr and K-Ar study of the granitoid batholiths of Peninsular Malaysia, proposed episodes of granitoid intrusion from late Carboniferous to late Triassic. They documented 230 Ma and 260 Ma events in the East Coast Province and a minor 200 Ma event in the Central Province (their Central Province and East Coast Province represent subdivisions of Stauffer's (1974) East Malaya Block, see Figs. 4, 5). 200 Ma, 230 Ma, 280 Ma and possible events >360 Ma were postulated for the West Coast Province. Ages obtained by both K-Ar mica and Rb-Sr whole rock methods are generally concordant for the East Coast Province. Ages yielded by both these methods are highly discordant for samples from the West Coast Province where K-Ar mica ages ranged from 40 to 210 Ma. Proposed intrusive ages for the West Coast Province were, consequently, based only on Rb-Sr whole rock isochrons. Where insufficient samples were analysed, apparent ages were calculated based on the assumption of an initial $^{87}$Sr/$^{86}$Sr ratio of $\sim$0.708.

Bignell and Snelling (1977) also noted the generally higher Sr initial ratios ($\sim$0.708-0.717) of the West Coast Province granitoids compared to those estimated for the East Coast Province granitoids ($\sim$0.707-0.708). Attention was also drawn to the contrast between the felsic, highly differentiated 'tin granites' of Peninsular Malaysia and the metaluminous, largely tonalitic cordilleran batholiths of North and South America.

Rajah et al. (1977) summarized the work done by geologists of the Geological Survey of Malaysia on the granitoids of the East Coast Province. They discussed the field and petrographic features of the granitoid masses, aspects of the major element chemistry and associated mineralization.
Fig. 4. Tectonic divisions of Peninsular Malaysia as described in Bignell and Snelling (1977). The medial suture represented by the Bentong-Raub 'ophiolite line' separates the West Coast Province from the Central and East Coast Provinces. See also Fig. 5.
Hutchison (1977) divided Peninsular Malaysia into four petrological-tectonic zones (Fig. 5) based on the chemical, mineralogical and isotopic characteristics of the granitoid bodies and their nature of emplacement. He divided the western block into a Western Stable Shelf and a Main Range Belt and the eastern block into a Central Intrusive Belt and an Eastern Belt.

Mitchell (1977) examined the regional geology of the Southeast Asian region and suggested tectonic settings for emplacement of the Southeast Asian 'tin granites'. In his model, Mitchell (1977) did not consider the possibility of pre-Triassic granitoids in the western block. He interpreted the granitoid batholiths of the eastern block as representing a plutonic arc related to eastward subduction along the Bentong-Raub line. Granitoids of the western block were interpreted to be generated during the late Triassic continental collision of the western and eastern blocks. Hutchison (1978) discussed these tectonic settings of the granitoid batholiths and examined the possibility of fitting pre-Triassic plutonic activity in the western block into a regional tectonic picture.

Hamilton (1979) reexamined the data of Bignell and Snelling (1977) and suggested that the old Rb-Sr whole rock isochron ages yielded by assuming a model initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of $\sim 0.708$ (Fig. 6) could also be interpreted as younger intrusions with high Sr initial ratios of around 0.72. He rejected radiometric apparent ages indicating pre-Triassic intrusions in the western block and expressed his preference for the granitoids in this block being of mainly Jurassic and possibly, Triassic age. Hamilton (1979) further utilised the data of Bignell and Snelling (1977) to show that present day $^{87}\text{Sr}/^{86}\text{Sr}$ values of the granitoids in the western block displayed
Fig. 5 Outline summary map of Peninsular Malaysia showing the granite distribution in relation to the tectonic pattern (after Hutchison, 1977).
Fig. 6 $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{87}\text{Rb}/^{86}\text{Sr}$ isochron diagram for whole rock samples of granitoids from Peninsular Malaysia. Model isochron ages are calculated based on an initial ratio of 0.708. Diagram is taken from Hamilton (1979) and is based on Hamilton's (1979) simplified re-examination of the data of Bignell and Snelling (1977).
a hyperbolic pattern in a plot against Sr (ppm) and a straight line in a plot against Rb/Sr ratios. He attributed these patterns to mixing between a crustal component with high Rb and high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and a mantle component with high Sr and low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. He, however, erred in doing this without correcting for $^{87}\text{Sr}$ caused by decay of $^{87}\text{Rb}$ and totally disregarded the possibility that the high Rb/Sr ratios can be simply explained by extensive fractionation.

Hamilton (1979, p.80) also interpreted the granitoids of the western block to be of mainly Jurassic age and related to Jurassic (?) - Cretaceous subduction along the Indoburman melange belt. It will be shown in chapters three and four that the granitoids of the western block are of Triassic age, not Jurassic or younger and that their characteristics are in contrast to those of Tertiary subduction-related batholiths.

Extensions of the granitoid terrains of Peninsular Malaysia into Thailand and Indonesia have been studied by a number of workers. These studies have allowed the correlation of the granitoid provinces of Southeast Asia to be made with more confidence. Published works on the Thai granitoids include Garson et al. (1975), Suensilpong et al. (1977), Pongsapich and Mahawat (1977), Beckinsale (1979) and Ishihara et al. (1980). Priem et al. (1975) and Jones et al. (1977) reported results of their studies of the granitoid bodies on the Indonesian islands of Bangka and Belitung.

A very recent development that may potentially be of great importance is the initiation of a collaborative study in 1981-82 between the Institute of Geological Sciences of London and the Geological Survey of Malaysia aimed at detailed mapping of the lithological variations within the granitoid batholiths of Peninsular
Malaysia. The initial results were discussed in a short note by Cobbing et al. (1982) and in a brief personal correspondence from Cobbing in 1982. Cobbing et al. showed that the batholiths, based on their preliminary data, could be subdivided into smaller intrusive units with abrupt intrusive contacts. They have been able to recognize more detailed sub-units than has been attempted in this thesis but their major units are broadly equivalent to the suites discussed in this thesis. I have not attempted any detailed division of the batholiths because the thrust of this study is primarily geochemical and isotopic.

Division of tectonic and plutonic provinces

For descriptive purposes, I have opted to retain the tectonic division of Peninsular Malaysia into West Coast, Central and East Coast Provinces (Fig.4) as used by Bignell and Snelling (1977). This nomenclature was retained to avoid uncertainties in the frequent references to these workers' data in this thesis. The relatively minor intrusive suites in the Central Province have, however, not been studied. The batholiths studied in this thesis are those in the East Coast and West Coast Provinces only. Based on the tectonic reconstruction discussed in an earlier section of this chapter, the granitoid batholiths of Peninsular Malaysia can be best described in terms of a late Palaeozoic - Triassic subduction related East Coast Province and a late Triassic continent collision related West Coast Province.
CHAPTER THREE

GEOCHRONOLOGY OF THE PENINSULAR MALAYSIAN BATHOLITHS

INTRODUCTION

The U-Pb dating method possesses a significant advantage over other radiometric methods in that dual decay of U to Pb ($^{238}\text{U} \rightarrow 206\text{Pb}$ and $^{235}\text{U} \rightarrow 207\text{Pb}$) allows two separate ages to be calculated using identical parent and daughter elements in a single mineral. Open system behaviour, resulting in loss or gain of either U or Pb, may be detected by a disagreement between $^{206}\text{Pb}/^{238}\text{U}$ or $^{207}\text{Pb}/^{235}\text{U}$ ages. Zircon is an excellent mineral for U-Pb dating. It crystallizes with high U and very low Pb contents initially, so that Pb measured in zircon analyses is almost entirely radiogenic and U-Pb ages calculated are not sensitive to any errors in common Pb correction. High temperatures required to reset U-Pb zircon systems facilitate the use of the method for determining the chronology of plutons in complex batholiths where resetting of K-Ar and Rb-Sr mineral ages commonly is a serious problem. Concordant zircon U-Pb ages (those where the $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ apparent ages are identical) are consequently very reliable estimates of the time of zircon crystallization in granitoid plutons, and the error of a single analysis (typically ± 2-3 Ma) can be assigned to such crystallization ages.

Inheritance and radiogenic Pb loss processes

In practice, the effects of inheritance (survival of old zircons in a younger rock) and radiogenic Pb loss may make zircon U-Pb analyses difficult to use for geochronological purposes. However, the effects of these processes can be recognized and sometimes
circumvented by plotting analyses on a concordia diagram (Wetherill, 1956). The concordia curve, which is the locus of present day $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ values for closed U-Pb systems of all ages is widely used as a reference in discussions of Pb loss (Silver and Deutsch, 1963) and inheritance phenomena (Grauert and Hofmann, 1973).

A number of processes have been proposed to explain the effects of Pb loss (see Gebauer and Grünenfelder (1979) for a summary). These include episodic disturbances (Wetherill, 1956), continuous diffusion (Nicolaysen, 1957), low temperature Pb loss after extensive time-integrated radiation damage and alteration (Krogh and Davis, 1974, 1975), low temperature annealing (Gebauer and Grünenfelder, 1979) or some combination of the various possible mechanisms. Pb loss yields points that plot below the concordia curve and calculated U-Pb ages that are younger than the true age of zircon crystallization. The discordance line defined by cogenetic zircons that have suffered different amounts of Pb loss may be extrapolated to the point representing zero Pb loss, namely its upper intersection with the concordia curve, to obtain the zircon crystallization age. Discordance lines with upper intersections that represent magmatic or high-temperature metamorphic ages are termed normal discordia. The interpretation of lower intersection ages depends upon the assumed cause of discordance. If Pb loss occurred during a single episodic disturbance (Wetherill, 1956), the intersection will date this episode. If Pb loss has occurred by diffusion (Nicolaysen, 1957) or a combination of processes, the lower intersection age will be geologically meaningless.
Zircon inheritance yields points that plot below the concordia curve and calculated U-Pb ages that are older than the true age of zircon crystallization. The discordance line defined by mixing a magmatic zircon component with an older inherited zircon component may be extrapolated to the point corresponding to zero inherited component, namely the lower intersection with the concordia curve, to yield an apparent age of the magmatic component. This apparent age will represent the true age of zircon crystallization only if no post-emplacement Pb loss has occurred and ideal mixing relationships between two simple zircon populations are valid. By the same reasoning, upper intersections with the concordia curve yield apparent ages that represent true ages of the older components only if these were concordant or lost Pb only during the event recorded by the lower intersection. In this study, the term reverse discordia is used to describe discordance lines below the concordia curve that result from the effects of zircon inheritance. This usage needs to be distinguished from another category of reverse discordant systems with excess radiogenic Pb that yield points plotting above the concordia curve (e.g. Rosholt et al. 1973).

The greater the loss of Pb or incorporation of old radiogenic Pb, the more discordant is the result. Ages inferred from concordia intersections of discordia chords defined by highly discordant samples have much greater uncertainties than the 2-3 Ma uncertainty of concordant data. Uncertainties due to extrapolation are, however, less important than those introduced by complex processes (e.g. combined Pb loss and inheritance) that yield non-linear discordance patterns. In such cases, crystallization age estimates are difficult.
Some U-Pb zircon geochronological case studies

A number of U-Pb zircon studies of Phanerozoic granitoid batholiths have been made in recent years. Among the youngest batholiths studied using the method are those of the western U.S.A. (e.g. Mattinson, 1978; Chen, 1982; Bickford et al., 1981). In general, the results of Mattinson (1978) for the Salinian Block plutons of California and Chen (1982) for the central Sierra Nevada Batholith indicate minimal zircon inheritance effects. The granitoids in these two regions are Mesozoic in age and show characteristics of the I-type granitoids of Chappell and White (1974). This general lack of significant zircon inheritance in I-type granitoids seems to be a widespread feature. Some of the zircon populations studied by Mattinson (1978) and Chen (1982) are both internally concordant (\(\frac{^{206}\text{Pb}}{^{238}\text{U}}\) and \(\frac{^{207}\text{Pb}}{^{235}\text{U}}\) ages are identical) and externally concordant (different fractions yield identically concordant U-Pb ages within analytical error). A number of plutons in the Sierra Nevada Batholith, however, possess zircon populations that are internally concordant but externally discordant. Such behaviour is attributed to Pb loss that shifts the original coordinates of a system down towards the origin along a straight line. Because the concordia resembles a straight line in the 0-200 Ma age range, Pb loss discordia chords overlap with concordia resulting in 'concordant' U-Pb points defining a discordia chord. A comparison (Chen, 1982) of the ages obtained from various radiometric clocks including U-Pb zircon, U-Pb sphene, K-Ar hornblende and K-Ar biotite indicated that U-Pb zircon ages, where Pb loss is not evident or can be evaluated to yield minimum age estimates, generally best approximate the emplacement age of a pluton. Mattinson (1978) established differences in cooling histories from different plutonic terrains in the Salinian Block by using U-Pb zircon
ages as emplacement ages and comparing younger K-Ar and fission-track ages with these emplacement ages.

In contrast to the Sierra Nevada and Salinian Block granitoids, zircon populations from the late Cretaceous-Tertiary Idaho Batholith located within the 'Inner Cordilleran S-type' belt of Miller and Bradfish (1981) display strong inheritance features. Lower intersections of zircon discordia chords (Bickford et al., 1981) are in the range 45-75 Ma and upper intersections are mainly in the range 1700-1800 Ma. The lower intersection ages are in broad agreement with K-Ar mineral ages (Armstrong et al., 1977; Criss et al., 1982). Although the interpretation that the zircon arrays had preserved their crystallization ages in the lower intersections of the discordia chords was preferred and is in accord with the intrusive relationships of the various plutons, Bickford et al., (1982) conceded that, for some arrays, the interpretation that the patterns resulted from mixing between old zircons of variable ages and discordant magmatic zircons that had lost Pb could not be dismissed. Some apparently concordant monazite ages younger than zircon lower intersection ages were obtained and could be the result of Pb loss caused by post-emplacement thermal events. It could not be fully ascertained whether such loss had affected the zircon discordia chords.

The U-Pb zircon study of the Lower Palaeozoic Berridale Batholith of southeastern Australia by Williams (1977) was fully integrated with detailed geological mapping, mineralogical, chemical and other isotopic studies (White et al., 1977; Williams et al., 1975; Chappell and White, 1974). Contrasting discordia patterns obtained correlated well with the separation into S- and I-type granitoid suites. Zircon populations from I-type plutons either showed minor
inheritance behaviour or defined normal discordant patterns. S-type plutons were shown to contain zircon populations that display marked inheritance patterns, yielding upper intercepts in the general age range 1600–1900 Ma. Pb loss was shown to have affected many of the inheritance patterns which showed poorly defined and geologically meaningless lower intersection ages younger than concordant monazite, Rb-Sr mica and K-Ar mica or hornblende ages. Details of coupled Pb loss and inheritance patterns were further documented by the high-resolution ion microprobe work of Compston and Williams (1982). Spot analyses on single zircon grains from a granodiorite pluton showed up concordant U-Pb regions around 430 Ma, regions that had lost ~15% Pb, a single core that had lost almost all radiogenic Pb, had high U (10,000 ppm) and high common \(^{206}\text{Pb} / {^{204}\text{Pb} \approx 100}\) and old zircon cores that displayed variable Pb loss and which defined discordia indicating \(~1800 \text{ Ma and } ~2800 \text{ Ma protoliths (Fig.1).}

The work of Pidgeon and Aftalion (1978) on the Caledonian plutons of Scotland and England highlighted the usefulness of the U-Pb zircon method as a geochronological and petrogenetic tool. The zircon results clearly separated the emplacement episodes of the Older Granites (mainly 470 to 570 Ma) from those of the Newer Granites (~400 Ma). It was also shown that the Older and Newer Granites north of the important Highland Boundary Fault (HBF) show zircon inheritance indicating Proterozoic protoliths. Newer Granites south of the HBF do not show evidence of the presence of inherited zircon. These results were interpreted to indicate that the HBF represented a fundamental structure separating a dominantly Proterozoic basement to the north of the fault from a young, Palaeozoic basement to the south.
Fig. 1A-B Ion microprobe data for zircons from the Dalgety Granodiorite showing features of combined Pb loss and inheritance patterns. Conventional analyses represented by solid dots do not show a consistent relationship that allows a single discordia chord to be fitted (after Compston and Williams, 1982).
To summarize the significance of these selected Phanerzoic case studies, the U-Pb zircon radiometric clock has proved to be an important dating tool in granitoid terrains. Like all radiometric clocks, U-Pb zircon results need to be examined in the context of the results obtained by other methods. The clock suffers from uncertainties caused by Pb loss but patterns resulting from such loss are quite easily recognizable. Where Pb loss is minor, resultant ages generally represent the best estimates of emplacement ages of plutons in complex batholiths. An added advantage of the method is the possibility of extracting granitoid source information from U-Pb zircon studies.

ANALYTICAL TECHNIQUES

U-Pb zircon chemistry and mass spectrometry are based on the methods described by Krogh (1973). Page (1981) discussed aspects of the techniques adopted in this laboratory. Details of procedures, analyses of standards, blanks and error estimates may be found in the appendix. Pb was loaded using the $\text{H}_3\text{PO}_4$ - silica gel technique and U was loaded using $\text{H}_3\text{PO}_4$ - Ta$_2$O$_5$ on single Re filaments. Pb analyses are corrected for a 0.3 ng processing blank. Uncertainties in $^{207}\text{Pb}/^{206}\text{Pb}$ and Pb/U ratios are estimated to be 0.3% and 1.0% respectively. Decay constants used are those of Steiger and Jäger (1977). Calculation of errors follow those suggested by Ludwig (1980). Table 1A-B is a list of all U-Pb zircon analytical data discussed in this chapter.
'1

Table lA

U-Pb zircon analytical data for the West Coast Provin c e plutons

sample ; size

weight

f r actio n (µ )

(mg )

13
13
13
13
14
14
14
14
14

29
29
29
35
35
35
35
38
38
38
39
39
39
39
41
41
41
45
45
45
45
78
78
78
78
78
78

+110
8.86
-110 + 45A
4.47
-110 + 45B
3.52
-45
0.63
+170 brown
7.08
+170
4.32
-170 + 110
7.59
-110 + 70 brown 4.62
-70 brown
5.27
+170
9.08
-70 + 45
6.14
-45
2.52
+170
0.54
-110 + 70
0.93
-70 + 45
1.86
-45
1.42
-110 + 70
7.89
-70 + 45
6.41
-45
4.40
+170
2.63
-170 + 110A
9.53
-170 + ll0B
7 .91
-110 + 70
6.52
-170 + 110
7.48
-110 + 70
6 . 02
-70
4 .11
-170 + 110
3.10
-110 + 70A
3.90
-110 + 70B
2.20
-70
6.08
-250 + 170A
11. 74
-250 + l70B
2. 72
-170 + 110
8.88
-110 + 70
7.33
5 .18
-70 + 45
-45
3.20

Table lB

132
132
132
132
120
120
120
120
106
106
106

u

Pb

( ppm )

(ppm)

1072.2
1064.2
1408.8
2476.7
2875.4
1135 .3
1092. 7
2552.8
2706.4
2173.9
2843.l
2934.3
1267.3
1463.7
1708.3
1843.6
4237.0
3808.4
3527.0
1843.3
1482.6
2416.8
1943.6
1764.6
2192 .5
2388.4
1973.8
5545.2
2249.l
5017 .9
669.2
1931.2
742.l
828.4
999.2
1014.5

44.91
4 7 .-22
56.65
114 .06
89.31
37.95
37.09
81.77
87.61
71.44
95.06
99.05
45 .13
50.70
59.35
60.98
145.85
126.74
113 .so
62 .11
52.70
80.31
65.96
57.58
71.05
80.22
74.74
194.90
80.09
180.13
26.25
69.42
29.99
31.26
36.50
38.02

34a

20G Pb (mea s)
2o" Pb

atom % radio geni c Pb *

1035.4
1521.3
922.9
4196
4009
1413
2148
2410
1022
8137
2261
1440
911.l
2684
3397
1981
418.2
561.6
796.5
3369
2487
7252
3443
3708
3929
1498
600.5
513.4
772.4
452.5
883.4
1030
688.8
4202
6095

84.18
84.85
84.55
67.54
89.32
85.92
86.28
88.82
86.24
90.33
89.30
88.41
86.24
88.27
88.89
88.82
81.28
83.73
85.42
89.80
88.34
90.81
89.55
86.42
86.19
83.74
81.55
83.33
84.18
82.33
80.78
84.25
80.73
85.67
85.95
85.81

3141

206

r - atomic r a t ios - - i r - apparent ages--,
201 Pb
201Pb
2 oGPb 201 Pb 201Pb 206 Pb

--

-23su

23au

.05741
.06225
.05710
.05440
.05110
.05134
.05230
.05060
.05155
.05209
.05112
.05131
.05298
.05395
.05455
.05178
.05084
.04975
.04991
.05178
.05398
.05279
.05245
.05003
.05005
.05070
.05222
.05061
.05172
.05024
.05192
.05145
.05392
.05386
.05391
.05463

.31870
.37142
.30526
.27066
.22650
.23356
.24368
.22920
.22694
.24749
.24279
.24253
.2582 4
.2638 9
.2697 5
.2423 6
.2176 5
.21478
.21477
.25025
.26983
.25489
.25435
.22522
.22315
.22582
.24863
.22888
.24277
.22808
.25818
.24595
.27448
.27811
.27085
.28038

.0402 6
.043 27
.03877
.03609
.032 15
.03300
.03 379
.03285
.03193
.0344 6
.03445
.03428
.03 535
.03547
. 03586
. 03 39 5
. 03 105
.031 31
.0312 1
.0350 5
.03625
.0350 2
.03517
.03265
.03233
.03230
.03455
.03280
.03404
.03292
.03607
.03467
.03692
.03745
.03644
.03722

.05324
.05262
.05196
.05209
.05074
.05106
.05123
.05122
.05256
.05814
.05161

.32693
.31703
.30029
.30221
.25101
.23617
.24426
.23255
.30961
.30051
.29693

.04453 339
.04369 313
.04192 284
.04207 290
.03588 229
.03355 243
.03458 251
.03293 251
.04272 310
.04205 278
.04172 268.4

207

208

206 Pb

5.981
6.017
6.065
3.687
4 .892
5.196
5.049
4.982
5.675
4.846
5.120
5.379
5.227
4.918
4.942
5.026
6.964
6.336
5.809
4 .924
5.257
4.955
5.037
4.620
4.605
5.022
6.628
6.576
5.856
6.782
5.480
5.330
5.996
4.846
4.756
4.974

9.759
9.081
9.295
28. 77
5. 772
8.834
8.631
6.164
8.006
4.819
5.545
6.149
8.491
6.799
6.159
6.129
11.561
9.782
8.665
5.258
6.371
4.228
5.387
8.938
9.181
11.184
11.662
9.934
9.862
10.705
13.654
10.355
13 .157
9.470
9.282
9.198

15.742
9.168
9.444
8.404
15.978
16.120
14.791
15.051
8.933
7.758
7.538

2 o r, Pb

23su

23au

507
683
495
388
245
256
299
222
265
290
246
255
328
369
394
276
234
184
191
27 6
370
320
305
197
198
227
296
223
273
207
282
261
368
365
367
397

281
321
271
243
207
213
221
209.5
208
225
221
220
233
238
242
220
199 . 9
197 . 4
197.6
22 7
24 3
231
230
206. 2
204 . 5
206 . 7
225
209.2
221
208.6
233
223
246
249
24 3
251

254
273
245
228
204
209
214
208.5
203
218
218
217
224
225
227
215
197.l
198.5
198 . 2
222
230
222
223 .
207 . l
205.l
204 .9
21 9
208. 0
216
208 .8
228
220
234
23 7
231
23 6

287
280
267
268
227.4
215
222
212
274
267
264.0

281
276
264
266
227.2
213
219
209
270
266
263. 5

U-Pb analytical data for zircons from East Coast Province plutons

+170
-170 + 110
-110 +70
-70
+170
-170 + 110A
-170 + ll0B
110
-110 + 70
70 + 45A
-70 + 45B

9 .12
10.79
9.03
5.61
2.89
9.58
12.95
7.83
8.60
4.69
4.60

1016.9
1177.7
1562.5
1831.5
1167 .1
1205.0
1239.2
1267.6
1721.5
3362.0
3457.3

48.81
51.69
65.96
76.82
48.99
47.02
48.08
47.07
74.19
140 .13
142.63

11316
3953
3843
3400
365.2
435.4
595.4
566.l
2232
2954
2802

79.93
86.00
85.78
86.71
77 .09
77 .20
79.10
78.82
85.97
87.28
87.50

4.324
4.812
4.753
4.859
6.736
6.500
5.976
5.999
5.061
4.931
4.937

* After correction for a total Pb blank of 0.3 ng ; isotopic composition of
la bora to r y acid blank= 206 Pb / 20 4Pb = 17.9 7; 207 Pb / 20 4Pb = 15.55 ; 2o ePb / 201+ Pb = 37 . 71


Table 2: Rb-Sr biotite and whole rock analyses

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>$^{87}\text{Rb}/^{86}\text{Sr}$</th>
<th>$^{87}\text{Sr}/^{86}\text{Sr}$</th>
<th>Biotite model age</th>
</tr>
</thead>
<tbody>
<tr>
<td>14 biotite</td>
<td>1462</td>
<td>7.63</td>
<td>647.9</td>
<td>2.4635</td>
<td>190 ± 2 Ma</td>
</tr>
<tr>
<td>35 biotite</td>
<td>1054</td>
<td>5.17</td>
<td>689.9</td>
<td>2.4568</td>
<td>177 ± 2 Ma</td>
</tr>
<tr>
<td>45 biotite</td>
<td>733</td>
<td>5.69</td>
<td>394.9</td>
<td>1.3408</td>
<td>111 ± 1 Ma</td>
</tr>
<tr>
<td>13 WR</td>
<td>409</td>
<td>62</td>
<td>19.25</td>
<td>0.79240</td>
<td>-</td>
</tr>
<tr>
<td>14 WR</td>
<td>306</td>
<td>74</td>
<td>12.02</td>
<td>0.76004</td>
<td>-</td>
</tr>
<tr>
<td>29 WR</td>
<td>423</td>
<td>40</td>
<td>31.17</td>
<td>0.82719</td>
<td>-</td>
</tr>
<tr>
<td>38 WR</td>
<td>643</td>
<td>30</td>
<td>64.20</td>
<td>0.93076</td>
<td>-</td>
</tr>
<tr>
<td>45 WR</td>
<td>319</td>
<td>51</td>
<td>18.05</td>
<td>0.78240</td>
<td>-</td>
</tr>
<tr>
<td>73 WR</td>
<td>353</td>
<td>67</td>
<td>15.25</td>
<td>0.77060</td>
<td>-</td>
</tr>
<tr>
<td>78 WR</td>
<td>254</td>
<td>146</td>
<td>5.02</td>
<td>0.74146</td>
<td>-</td>
</tr>
<tr>
<td>106 WR</td>
<td>277</td>
<td>31</td>
<td>25.65</td>
<td>0.81187</td>
<td>-</td>
</tr>
<tr>
<td>111 WR</td>
<td>281</td>
<td>74</td>
<td>11.04</td>
<td>0.75455</td>
<td>-</td>
</tr>
<tr>
<td>120 WR</td>
<td>166</td>
<td>158</td>
<td>3.03</td>
<td>0.71708</td>
<td>-</td>
</tr>
</tbody>
</table>

In subsequent diagrams, sample numbers of these new analyses are marked by asterisks to distinguish them from data of Bignell and Snelling (1977).

* Biotite model ages calculated based on initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.720$
GEOLOGICAL BACKGROUND AND PREVIOUS WORK

The main objective of this chapter is to present the results of a U-Pb zircon study of selected granitoid plutons and to discuss these results in the context of previous geochronological work. A subsidiary objective is to examine the implications of the U-Pb zircon data in relation to processes of granitoid genesis.

It became obvious from the works published in the 1970's (Stauffer, 1974; Bignell and Snelling, 1977; Hutchison, 1977; Mitchell, 1977), that Peninsular Malaysia consists of two distinct geological blocks (Fig.2) separated by a fundamental medial-trending structural break. Because this break is defined by a belt of rocks featuring the association of serpentinites and cherts with materials resembling melange, Hutchison (1975) described the break as the Bentong-Raub ophiolite line. Although actual plate tectonic reconstructions differ in detail, the consensus is that the Bentong-Raub ophiolite line represents a tectonic suture joining two different crustal blocks. Hutchison (1977) summarized the data available up to 1977 and described the contrasting mineralogical, chemical, isotopic and emplacement characteristics of the granitoid batholiths on either side of the suture. Geochronological studies to document the ages of the granitoid batholiths in Peninsular Malaysia were initiated by the Institute of Geological Sciences (IGS) of London in 1964 and the results of the Rb-Sr and K-Ar work were published in 1977 (Bignell and Snelling, 1977). The K-Ar data and much of the Rb-Sr data discussed hereafter are from this important study. Following Bignell and Snelling (1977), the two main plutonic provinces of Peninsular Malaysia are called the West Coast and East Coast Provinces (Fig.2).
Fig. 2. Distribution of granitoid batholiths in the West Coast and East Coast Provinces of Peninsular Malaysia and locations of U-Pb zircon samples.
The West Coast Province

A striking feature of the dating results of the West Coast Province is the large discrepancies between K-Ar mica ages and Rb-Sr whole rock isochron ages. A histogram of the K-Ar ages (Fig.3) shows a range from 35 to 210 Ma, whereas regional Rb-Sr isochrons were interpreted by the IGS workers to indicate three main intrusive episodes – late Carboniferous (~280 Ma), early Triassic (~230 Ma) and late Triassic (~200 Ma). Minor Devonian (~360 Ma) and Silurian (~430 Ma) intrusive episodes were tentatively suggested, based on Rb-Sr analyses of a number of widely scattered samples. The absence of K-Ar ages older than 210 Ma was interpreted to be the result of Ar loss caused by the late Triassic intrusions and younger fault-related disturbances. The presence of granitoids older than Triassic was disputed by some workers. Hamilton (1979), for example, questioned the assumption that Rb-Sr data for reconnaissance samples from a limited geographic region plotting along a linear array from an initial ratio of ~0.708 - 0.710, probably defined isochrons. He suggested that some of the granitoids are of Triassic age but most are of Jurassic age and characterised by high initial ratios. Zircon U-Pb work is an independent means of testing the differing interpretations. The present U-Pb zircon work has, in addition, benefited from field mapping and a mineralogical and chemical characterization that have assisted in constraining the geochronological interpretations.

The East Coast Province

K-Ar mineral (biotite and hornblende) ages range from 190 to 270 Ma for plutons of the East Coast Province. A small number of determinations yield disturbed ages as young as 110 Ma. Because of the restricted scope of the IGS study, interpretation of the Rb-Sr
Fig. 3. Histogram of K-Ar ages from the West Coast Province (after Bignell and Snelling, 1977).
data rests heavily on regional Rb-Sr isochrons that suggest intrusive episodes at \( \sim 260 \) Ma and \( \sim 230 \) Ma. Although lacking in detail, the Rb-Sr and K-Ar results are compatible with each other and with stratigraphic constraints. The present U-Pb zircon work is an attempt to independently date some of the same plutons which geochronologically are of most interest.

**U-Pb Zircon Dating of the West Coast Province Plutons**

(A) Penang Island

Granitoid plutons exposed on Penang Island (Fig.4) represent the westernmost outcrops of the West Coast Province S-type suites in Peninsular Malaysia that were studied (Fig.2). The plutons are biotite adamellites and granites with variable muscovite and tourmaline. Field mapping by students from the University of Malaya was summarized by Ng (1978) and reproduced in Fig.4. Four distinct plutons could be distinguished although each is petrographically variable in detail. Their suggested formal names are shown in Fig.4. Intrusive relationships have not been worked out precisely due to limited exposure in critical areas and complications arising from irregular patches of pegmatitic and aplitic phases. The shapes of the plutons suggest that the northern Bt. Bendera Granite is intrusive into the central Sg. Ara Granite, which itself intrudes the Gertak Sanggul Adamellite. This interpretation is not unequivocal, however, because fault contacts cannot be ruled out. How the Gelugor adamellite fits into this sequence could not be worked out as it is in fault contact with the other plutons.
Fig. 4. Distribution of granitoid plutons on Penang Island (simplified from Ng, 1978). Numbers beside solid dots represent either K-Ar mica ages (after Bignell and Snelling, 1977) or new Rb-Sr biotite ages.
Fig. 5A. Plot of granitoids of Penang Island on a modified ACF diagram (White and Chappell, 1977)

Fig. 5B-E. Selected Harker variation plots of the Penang Island granitoids.
Symbols as in Fig. 5A.
Geochemical data support the field division of plutons. Figs. 5A-E show the chemical separation of the various plutons. Notable chemical differences are the highly peraluminous nature, high $P_2O_5$, Nb and low Zr, Y abundances of the Sg. Ara Granite that serve to distinguish it from the Bt. Bendera and Gelugor masses. Consistent chemical trends displayed by the Bt. Bendera and Gelugor plutons suggest that they represent members of a single fractionation suite.

Six IGS K-Ar mica ages and a new Rb-Sr biotite age (sample 14, Table 2) are displayed in Fig. 4. With the exception of a 68 Ma age from the Bt. Bendera Granite in the vicinity of the main north-south trending fault, all mica ages are late Triassic - early Jurassic (185-205 Ma). Rb-Sr whole rock points display considerable scatter on an isochron diagram, even for samples within the same pluton (Fig. 6). The interpretation favoured by the IGS workers was that at least two, and probably three, periods of emplacement were indicated by the whole rock data. These were: ~290 Ma (Sg. Ara Granite, Gertak Sanggul Adamellite), ~200 Ma (Bt. Bendera Granite) and possibly an episode between 220-245 Ma (Gelugor Granite). The suggested late Triassic Rb-Sr isochron age for Bt. Bendera Granite is compatible with the oldest K-Ar biotite age of $205 \pm 6$ Ma obtained for this pluton but complete resetting needs to be invoked to explain the absence of K-Ar ages older than 205 Ma. Samples were collected from the Sg. Ara Granite and the Gelugor Adamellite for U-Pb zircon studies in order to test whether these plutons are in fact older than 220 Ma.

Zircons extracted from these granitoids display a number of characteristics that may be regarded as typical for the West Coast Province samples. Prismatic crystals, of fragments of such, having
Fig. 6. Rb-Sr isochron diagram of whole rock analyses for Penang Island granitoids (modified from Bignell and Snelling, 1977). Asterisks denote new analyses.

Fig. 7. Concordia diagram showing U-Pb zircon data for Sg. Ara Granite (sample 13) and Gelugor Adamellite (sample 14). Numbers beside zircon analyses of sample 14 represent concentration of U in ppm.
simple pyramidal terminations and highly variable length:width ratios of 1:1 to 10:1 are present in a single sample population. Colours are also highly variable and range from clear pink to opaque brown. Only for the former could any internal structure be discerned. Needle-shaped inclusions are commonly observed in such transparent pink grains but prominent altered cores such as those described by Bickford et al., (1981) were not observed.

The U-Pb results for the Sg. Ara Granite (sample 13) and the Gelugor Adamellite (sample 14) are shown on a concordia plot in Fig.7. Both zircon populations display reverse discordant patterns that may be attributed to the presence of old inherited zircon components. Intercepts with concordia of a best fit line to the four zircon points for sample 13 yield ages of $215^{+5}_{-6}$ Ma and $1560^{+150}_{-150}$ Ma. A conventional interpretation of the discordia is that the lower intercept of 215 Ma represents the time of zircon crystallization. Two IGS muscovite K-Ar ages of $194^{+5}_{-6}$ Ma and $185^{+4}_{-4}$ Ma are compatible with this interpretation if these K-Ar ages are taken to represent primary cooling ages. The results are not precise enough to uniquely separate the time of zircon crystallization at magmatic temperatures recorded by the zircon system of $215^{+5}_{-6}$ Ma and the time of the closure of the muscovite K-Ar system at $\sim 350^\circ$C (e.g. Harrison and McDougall, 1980). The best interpretation of the available data is that the Sg. Ara Granite was emplaced sometime between 195 Ma and 215 Ma. The fact that the zircon U-Pb age is similar to, but slightly older than the K-Ar muscovite age suggests that the zircon system has not suffered recent Pb loss. There is, therefore, reason to believe that the reverse discordia is generated primarily by various mixtures of old inherited zircon components, possibly with a mean age $\sim 1600$ Ma, and a 215 Ma magmatic zircon component.
U-Pb patterns displayed by zircons from sample 14 (Gelugor Adamellite) are more complex than those of sample 13. Both Pb loss and inheritance features can be recognized. Brown zircon fractions possessing high U concentrations (2580 - 2875 ppm) display a pattern indicative of variable Pb loss (Fig.7). A single brown fraction with the lowest U content (2550 ppm) plots within analytical error of concordia to yield $^{206}\text{Pb} / ^{238}\text{U}$ and $^{207}\text{Pb} / ^{235}\text{U}$ ages of 208.5 Ma and 209.5 Ma respectively. The three high-U brown zircon fractions are not sufficiently dispersed to define a reliable Pb loss chord. Pink zircon populations from sample 14 have considerably lower U contents (~1100 ppm) compared to the brown fractions. These pink zircon fractions define an apparent reverse discordance pattern. A discordance line fitted to the pink zircon points and the single internally concordant brown zircon point yields a linear array (MSWD = 1.5) with an upper intersection of 1700 Ma with a large uncertainty of ±600 Ma that primarily reflects the very limited dispersion of the points. The geological significance of the 209 Ma and 1700 Ma ages is dependent on the assumed nature of Pb loss and whether the pink zircon fractions may be treated as a separate population that has not suffered Pb loss. A useful constraint however, on any interpretation of the U-Pb data is that the U-Pb zircon crystallization age must be greater than the Rb-Sr biotite cooling age of 190 ± 2 Ma (sample 14, Table 2).

The gross U-Pb pattern (Fig.7) is suggestive of mixing between three components - a concordant magmatic component with a crystallization age around 210 Ma, a discordant magmatic component that suffered recent Pb loss caused by radiation damage, and inherited components that lost Pb during the ~210 Ma event (cf. Compston and Williams, 1982 and Fig.1).
points and the 2552 ppm U, internally concordant brown zircon point suggests that these zircon fractions may not have suffered post-emplacement Pb loss. In this case, the 209 Ma age of the internally concordant fraction may be taken as the age of magmatic zircon crystallization. A more complex scenario is possible. If all the fractions have suffered recent Pb loss, the discordance line generated may be regarded as fortuitous. For this case, the 209 Ma age of the 2552 ppm U zircon fraction represents a minimum age. Assuming a simple recent Pb loss model, this fraction yields an older limit of 222 Ma for the zircon crystallization age. The complexities of the pattern and the model-dependent nature of any interpretation does not allow an unequivocal magmatic crystallization age to be assigned. It is suggested that an emplacement age sometime between 209 Ma (apparently concordant U-Pb ages) and 222 Ma (207\(^{7}\) Pb/206\(^{6}\) Pb age of the 2552 ppm U brown zircon fraction) be provisionally accepted till more data becomes available. Such an age range is compatible with the 198 to 220 Ma ages obtained by U-Pb zircon dating for other West Coast Province S-type plutons (following sections).

To summarize, U-Pb zircon data presented above argue against the presence of granitoids older than 222 Ma on Penang Island. The difficulty in resetting zircon ages and the slightly older U-Pb zircon ages compared to younger K-Ar and Rb-Sr mica cooling ages suggest that the late Triassic U-Pb zircon ages (215 Ma for the Sg. Ara Granite, between 209 and 222 Ma for the Gelugor Adamellite) are approximating emplacement ages of the plutons studied. A reconciliation of these mineral ages and most of the Rb-Sr whole rock data is not difficult. Previous studies have shown the hazards of overreliance on Rb-Sr whole rock isochrons to obtain emplacement ages of granitoid plutons. Brooks and Compston (1965) documented high and nonuniform Sr initial
ratios for the S-type Heemskirk Granite of Tasmania and it is believed that such nonuniformity is the main cause of most of the scatter in the whole rock Rb-Sr plot of Fig. 6. The apparent good linearity of the 294 Ma best-fit line for the Sg. Ara Granite is, however, rather enigmatic. Roddick and Compston (1977) documented anomalously old 490 Ma Rb-Sr whole rock ages for the ~425 Ma Murrumbidgee Batholith of southeastern Australia and described these as 'inherited isochrons'.

McCarthy and Cawthorn (1978) discussed the importance of changes in initial Sr ratios by in-situ decay of $^{87}$Rb during protracted fractionation that has the effect of yielding apparent old ages. The latter factor is significant in view of the very high Rb/Sr ratios in many of the West Coast Province S-type suites and the fact that isochrons are often strongly controlled by high Rb/Sr points. By itself, however, the model of in situ increase of Sr initial ratio of a magma fails to explain the data because of the large radiogenic increase from an apparent initial ratio of ~0.725 for the lowest Rb/Sr point to ~0.745 for the highest Rb/Sr point. This large increase requires an unrealistically long fractionation history (>10 Ma which is the upper limit indicated by U-Pb zircon, K-Ar mica and Rb-Sr mica ages) and Rb-Sr ratios higher than those observed. On this basis, it is suspected that the Sg. Ara Granite may comprise several phases, each with high and variable initial ratios (thus, invalidating a single whole rock isochron) or that the apparent isochron represents an 'inherited isochron' of the type described by Roddick and Compston (1977). One geologically plausible way to generate 'inherited isochrons' is assimilation of wall rocks with highly radiogenic Sr. Assimilation can produce anomalously old apparent ages because of the slope generated on a $^{87}$Sr/$^{86}$Sr - $^{87}$Rb/$^{86}$Sr plot even before the whole rock system starts recording the in situ decay of
Lastly, the U-Pb zircon data for the two plutons indicate that mid-Proterozoic components, with probable ages of \(\sim 1600 - 1700\) Ma, have been incorporated in the granitoid magmas. These zircon inheritance ages for the plutons from Penang Island will be considered with those of the Main Range Batholith in a later section.

(B) The Kuala Lumpur region

The Kuala Lumpur region comprises that part of the Main Range Batholith from 3° 00'N to 3° 45'N (Fig. 8). The granitoids are intrusive into metasedimentary rocks of Silurian-Devonian, and possibly Carboniferous age (Jones, 1973; Stauffer, 1973). Rock types range from mafic granodiorite to leucogranite. Muscovite is commonly present as is tourmaline. Mn-rich garnet is developed in highly siliceous, highly differentiated aplitic bodies. The batholith in this region is divided into major segments by screens of country rocks. A central southeast-trending, in places fault-bounded, inlier of country rocks separates the batholith into an eastern and a western segment. Another elongated sedimentary screen to the north isolates a distinct granodiorite pluton in the northeastern corner of the study region from the other plutons of the eastern segment. This granodiorite pluton is referred to in this study as the Tranum Granodiorite. Limited exposures have prevented detailed mapping being carried out to distinguish other individual plutons within the segments.

Geochemical evidence indicates that the structurally separate batholith segments discussed above are each characterized by distinct chemical patterns. These segments are consequently interpreted to be
Fig. 8 Distribution of granitoids in the Kuala Lumpur region of the Main Range Batholith. Numbers in parentheses represent K-Ar mica ages (Bignell and Snelling, 1977) or Rb-Sr mica ages. Incorporates data from Shu (1970), Gobbett and Tjia (1973).
separate intrusive suites. The eastern suite is referred to in this study as the Ulu Kali suite and the western suite as the Beranang suite. In Figs. 9 A-C, selected Harker variation plots are used to show the distinct chemical fields of the Ulu Kali and Beranang suites and the separation of the field of the Tranum Granodiorite from that of the adjacent Ulu Kali suite. Plutons of the Beranang suite and the Tranum Granodiorite are characterized by low Y, La and Th compared to the Ulu Kali suite, for similar SiO₂ contents. Although these chemical differences may be reflecting source rock differences, they could also be the result of differences in fractionation histories. In particular, low La, Th abundances in high silica granites may be explained by early monazite fractionation in the Beranang suite and Tranum Granodiorite.

(1) Geochronology of the Beranang suite

The six IGS K-Ar mica ages reported range from 131 to 203 Ma (Bignell and Snelling, 1977). Two intrusive episodes were suggested by the IGS workers based on Rb-Sr whole rock work (Fig. 11). These episodes were believed to be dated as 245 Ma and 206 Ma. Very high $^{87}\text{Rb}/^{86}\text{Sr}$ (100-300) samples yield ~200 Ma ages whereas lower $^{87}\text{Rb}/^{86}\text{Sr}$ samples were shown as defining the 245 Ma isochron. However, field separation of the postulated older and younger granitoids is not possible and chemical data suggest a single chemical suite. Three localities were selected for U-Pb zircon work in an attempt to reconcile the conflicting field, chemical and geochronological inferences. Sample 29 is a muscovite-biotite leucogranite belonging to the IGS 245 Ma suite and sample 39 is a petrographically similar muscovite-biotite granite belonging to the IGS 206 Ma suite. Both samples were collected from the spatially
Fig. 9A-C Selected Harker variation diagrams illustrating chemical fields of the two main suites in the Kuala Lumpur region of the Main Range Batholith.
Fig. 10 U-Pb zircon data for plutons of the Beranang suite displayed on a concordia diagram.

Fig. 11 Rb-Sr data for whole rock samples from the Beranang suite and nearby plutons showing apparent isochrons if an initial ratio of \( \frac{87}{86} \)Sr is assumed. Modified from Bignell and Snelling (1977).
separate western lobe of the Beranang suite. Sample 35 is a mafic
grandodiorite sample from a previously undated locality within the
main Beranang mass.

U-Pb zircon patterns of all three samples indicate incorporation
of old zircon components. The zircon populations of sample 29 show
the least effects of inheritance and they form a linear array with the
zircon populations of sample 39 on a concordia diagram (Fig.10).
Combining the data sets of both samples yields a reverse discordia
chord with significant scatter (MSWD = 3.5) which intersects concordia
at $215^{+2}_{-4}$ Ma and $1650 \pm 400$ Ma. The lower intercept age of $215^{+2}_{-4}$ Ma
is interpreted as the time of zircon crystallization. This is
consistent with the $203 \pm 5$ Ma biotite K-Ar cooling age obtained by
the IGS for a sample from the same quarry as sample 29. The present
U-Pb zircon data are also consistent with petrographic and chemical
considerations that suggest that samples 29 and 39 are cogenetic. In
the light of these results, it is proposed that the emplacement age of
the pluton constituting the western lobe of the Beranang suite can be
confidently bracketed between $203 \pm 5$ Ma and $215^{+2}_{-4}$ Ma.

Zircon fractions from the biotite granodiorite sample 35 also
define a reverse discordia (Fig.10) with intercepts at $211^{+5}_{-8}$ Ma and
$1610 \pm 450$ Ma. The magmatic zircon crystallization age of $211^{+5}_{-8}$ Ma is
virtually indistinguishable from the $215^{+2}_{-4}$ Ma determined for the
pluton constituting the western lobe. This 211 Ma age is probably the
best estimate of the emplacement age of the granodiorite pluton
outcropping in the northeastern lobe of the Beranang suite. A
considerably younger age of $177^{+2}_{-4}$ Ma (Table 2) is recorded by the
Rb-Sr biotite clock. The large time difference between the Rb-Sr
biotite age and the U-Pb zircon age is indicative that the former
represents a disturbed, and not a primary cooling age.

In summary, the late Triassic (211 - 215 Ma) U-Pb zircon ages obtained from three localities within the Beranang suite are identical within analytical error. The chemical similarities, structural continuity and age equality all indicate that the Beranang suite represents a single coeval and cogenetic suite of plutons. The U-Pb zircon ages are compatible with the 206 Ma age indicated by the IGS Rb-Sr whole rock work but do not support the suggestion that granitoids older than 220 Ma are present. The consistency of the zircon ages, their compatibility with the oldest K-Ar mineral ages and the equivocal nature of the reconnaissance Rb-Sr whole rock isochrons lend confidence to the proposed late Triassic emplacement ages of the plutons. Acceptance of these ages implies that the Beranang suite is characterized by high and variable initial Sr ratios (0.720 - 0.730) or exhibits open system behaviour that would preclude any precise age determination by the Rb-Sr whole rock method.

(ii) Geochronology of the Ulu Kali suite

This southeast trending suite is bounded to the west by the Bt. Tinggi Fault zone and partially separated from the Tranum Granodiorite to the east by a long sedimentary screen. Rock types present are very variable texturally although the most common variety is a coarse-grained biotite adamellite - granite with megacrysts of K-feldspar. K-Ar biotite ages in this segment of the Main Range batholith range from 92 to 177 Ma. A comprehensive sampling effort along a northern transect was carried out by the IGS workers. The samples from this transect showed a wide range in \( \frac{Rb}{Sr} \) (10 to 159) and yielded a precise whole rock isochron age of 206 ± 2 Ma, \( Sr_i = 0.711 \) (Fig.12). Widely scattered samples from the southern
Fig. 12. Rb-Sr whole rock isochron diagram for samples from the Ulu Kali suite (data of Bignell and Snelling, 1977).

Fig. 13. U-Pb concordia diagram for zircons of samples 38 and 41 from the Ulu Kali suite.
portion of the Ulu Kali suite were suggested to be coeval with the emplacement of the plutons to the north. One sample from the eastern flank, however, plots above the 206 Ma isochron and was assigned to a 270-290 Ma suite together with the Tranum Granodiorite discussed in a subsequent section.

Two localities were selected for U-Pb zircon dating. Sample 41 is from the northwestern edge of the suite where the Rb-Sr work indicates a 206 Ma intrusive event; sample 38 is from the locality where ~280 Ma plutons were suspected. The U-Pb zircon results of both samples are displayed on a concordia diagram in Fig.13. Chronological interpretation of the results is straightforward. Three different fractions from each sample show U-Pb ages that are both internally and externally concordant i.e. different fractions have both U-Pb ages agreeing with one another within analytical error. $^{207}\text{Pb}/^{206}\text{Pb}$ ages display larger differences (Table 1A) but, in view of the large uncertainties involved in calculating Pb-Pb ages in the 0-200 Ma linear portion of the concordia curve (+ 8 Ma for a 0.3% uncertainty in $^{207}\text{Pb}/^{206}\text{Pb}$ measurement, ± 12 Ma for a 0.5% uncertainty), little emphasis will be given to such ages. The best estimate of the zircon crystallization age is probably that of the mean of the three $^{206}\text{Pb}/^{238}\text{U}$ ages for each sample as such ages are least sensitive to errors in common Pb corrections. The mean $^{206}\text{Pb}/^{238}\text{U}$ age for sample 41 is $206 \pm 2$ Ma and that for sample 38 is $198 \pm 2$ Ma. Late Triassic ages are again obtained. The U-Pb zircon age for sample 41 is identical to that of the IGS Rb-Sr isochron age of $206 \pm 2$ Ma but the suspected 280 Ma event is not supported by the zircon data for sample 38.
The difference between the zircon ages for the two Ulu Kali suite samples is beyond that which may be attributed to analytical uncertainty. On this basis, it is suspected that the Ulu Kali suite consists of plutons with different emplacement ages i.e. an extended intrusive history spanning an interval of \( \sim 10 \) Ma is indicated. Alternatively, sample 38 may belong to a different and younger suite. More detailed field mapping and chemical studies are required to distinguish the two possibilities.

(iii) Tranum Granodiorite

The emplacement age of the rather uniform, coarse-grained porphyritic biotite granodiorite pluton outcropping to the east of the sedimentary screen in the northern portion of the study region has been particularly problematic. K-Ar biotite determinations by the IGS yield cooling/reset ages of \( 160 \pm 3 \) Ma and \( 165 \pm 4 \) Ma. A new Rb-Sr biotite determination also yields a reset age of \( 111 \pm 1 \) Ma. Whole rock samples display a limited range in \( \frac{87}{86} \)Rb/\( \frac{87}{86} \)Sr (Fig. 14) and plot above the 206 Ma isochron obtained for the plutons to the immediate southwest that belong to the Ulu Kali suite. A Rb-Sr whole rock isochron age of \( 269 \pm 14 \) Ma was proposed by the IGS workers who, nevertheless, cautioned against an unquestioning acceptance of the age because "the very limited range in \( \frac{87}{86} \)Rb/\( \frac{87}{86} \)Sr (8.5 to 12.7) ratios renders any isochron calculation suspect". The isochron included a sample from the same locality as sample 38 where a \( 198 \pm 2 \) Ma U-Pb zircon age was demonstrated in the previous section. It is felt that this sample should be excluded from the isochron calculation. Exclusion of this sample (IGS sample 137, p. 32, Bignell and Snelling, 1977) and inclusion of a new sample from which zircons were separated for U-Pb work yields a model l isochron age of \( 235 \pm 30 \) Ma, \( \text{Sr}_i = 0.722 \).
Fig. 14. Rb-Sr whole rock isochron diagram for the Tranum Granodiorite samples with 236 Ma and 269 Ma reference isochrons shown, as discussed in the text. Incorporates data from Bignell and Snelling (1977).

Fig. 15. U-Pb concordia diagram for Tranum Granodiorite zircon data.
U-Pb zircon results (sample 45, Fig.15) indicate variable incorporation of old zircon components. Two fractions yield identical concordant ages of $208 \pm 2$ Ma whereas two other fractions display inheritance behaviour. A discordia chord fitted to the four data points yields an upper intersection age of $1240 \pm 300$ Ma. The results indicate zircon crystallization of $208 \pm 2$ Ma and incorporation of inherited zircon components with a mean age of $\sim 1250$ Ma. The zircon crystallization age of $208$ Ma indicated by the two concordant fractions is older than Rb-Sr and K-Ar biotite ages and younger than the Rb-Sr whole rock isochron age. The relevance of the mica ages is considered to be uncertain in view of the widespread resetting of such ages in the region. The U-Pb zircon age interpretation is, however, less ambiguous, particularly as it is consistent with the other late Triassic ages of the adjacent plutons in this portion of the Main Range batholith. It therefore is suggested that this U-Pb zircon age of $208 \pm 2$ Ma be accepted as the most reliable estimate of the emplacement age of the Tranum granodiorite. The Rb-Sr whole rock isochron age probably represents an 'inherited isochron' (Roddick and Compston, 1977) age with doubtful age significance.

(iv) Genting Sampah hypersthene microgranodiorite

A suite of volcanic and subvolcanic rocks is found associated with the mid-Palaeozoic metasedimentary inliers that separate the eastern Ulu Kali suite from the western Beranang suite. Accounts of the rhyolitic units and quartz porphyries associated with the metasediments are given in Hutchison (1973b), Shu (1969) and Haile et al. (1977). The quartz porphyries are clearly related to the rhyolitic units in the Genting Sempah area, and in this study these
rocks are called hypersthene microgranodiorite as related rock units appear to be intrusive into metasediments further to the south (Shu, 1969). These rocks are equivalent to the pyroxene-bearing labradorite 'granite' of Bignell and Snelling (1977). Because of their field association with mid-Palaeozoic metasediments that are, in part, of Devonian age (Haile et al., 1977), these rocks are generally considered to be of Palaeozoic age and indicative of Palaeozoic igneous activity in the West Coast Province.

A number of IGS K-Ar biotite age determinations, however, showed ages ranging from 175 to 211 Ma. The restricted range in $^{87}\text{Rb}/^{86}\text{Sr} (4.4 - 5.3)$ for the microgranodiorite precludes any reliable Rb-Sr whole rock isochron calculation. Assuming an initial ratio of 0.710, Bignell and Snelling (1977) tentatively suggested a ~430 Ma age. The possibility that "these distinctive granitic rocks could be members of a late Palaeozoic intrusive suite characterized by abnormally high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of about 0.72" (Bignell and Snelling, 1977) was held open. If, however, the hypersthene microgranodiorite is regarded as cogenetic with the rhyolites as is done so in this study and the Rb-Sr data sets of both units are pooled together, a model 1 isochron yields an age of $204 \pm 13$ Ma, $\text{Sr}_i = 0.726 \pm 0.002$. (Fig.16). The large uncertainty in the age estimate indicates this is not an entirely satisfactory result and the scatter of the whole rock points suggests a degree of open system behaviour. This late Triassic age, however, is compatible with K-Ar biotite ages (175-211 Ma).

U-Pb zircon results for the microgranodiorite (sample 78) are shown in Fig.17. The reverse discordia obtained is indicative of the presence of inherited zircon components. A discordia chord (MSWD =
Fig. 16. Rb-Sr whole rock isochron diagram for the Genting Sempah rhyolites and hypersthene microgranodiorite. Modified from Bignell (1972), Bignell and Snelling (1977).

Fig. 17. U-Pb concordia diagram for the Genting Sempah hypersthene microgranodiorite zircon data.
fitted to the analytical data intersects concordia at $219^{+5}_{-9}$ Ma and $1550 \pm 300$ Ma. The lower intercept age of $219^{+5}_{-9}$ Ma is interpreted to represent the zircon crystallization age. This age is compatible with the oldest K-Ar biotite age of $211 \pm 4$ Ma and the Rb-Sr whole rock isochron age of $204 \pm 13$ Ma discussed earlier. It is suggested that the intrusive-extrusive age of the volcanic and subvolcanic rocks of the Genting Sempah area can be bracketed within the age interval 211 to 219 Ma.

The 211-219 Ma age range is identical to that of the Beranang suite plutons to the west of the central metasedimentary screen. The indicated high Sr initial ratio of $0.726 \pm 0.002$ is also typical of the granitoid suites in the study region. As such, the present evidence supports the contention that the volcanic and subvolcanic rocks of the Genting Sempah area represent coeval eruptive and high-level intrusive equivalents of some of the granitoids of the Main Range batholith.

SYNTHESIS OF THE U-PB ZIRCON RESULTS FROM THE WEST COAST PROVINCE

The major finding of the U-Pb zircon study is the consistent and unambiguous late Triassic (198 – 220 Ma) emplacement ages of the granitoid plutons studied. These results are consistent with the interpretation that the oldest IGS K-Ar ages ($\sim 200 - 210$ Ma) are approaching emplacement ages, but are inconsistent with suggestions that granitoid intrusive events $> 220$ Ma are indicated by regional Rb-Sr whole rock isochrons. The U-Pb zircon data argue against the existence of granitoids older than 220 Ma in the two areas (Penang Island and the Kuala Lumpur region of the Main Range batholith) where Rb-Sr evidence for such granitoids were most confidently suspected. It is suggested that where Rb-Sr whole rock isochron ages older than
220 Ma have been proposed, these should be regarded as accidental—resulting from high and variable Sr initial ratios and, possibly, the existence of open system behaviour. Some of the older isochrons (e.g. Tranum Granodiorite, Sg. Ara Granite), however, represent 'inherited isochrons' that may be the result of assimilation of wall rocks with highly radiogenic Sr. It is clear that regional dating using the Rb-Sr whole rock isochron method may yield misleading ages with no geological significance when applied to granitoid rocks like the West Coast Province S-type suites (cf. Brooks and Compston, 1965; Vidal et al., 1982). Rb-Sr isochron ages should be treated with caution unless supported by ages measured by other radiometric techniques.

A number of plutons (Penang Island, Beranang suite) display U-Pb zircon ages that exceed the oldest recorded K-Ar and Rb-Sr mica ages by ~10 - 20 Ma. It cannot be ascertained as yet whether this 10 - 20 Ma gap represents the typical cooling times of the West Coast Province plutons or reflects rapid uplift that only occurred throughout the province at ~190 to 200 Ma.

Volcanic and subvolcanic rocks in the Genting Sempah area northeast of Kuala Lumpur do not represent expressions of Palaeozoic igneous activity as previously believed. A microgranodiorite unit yields a late Triassic U-Pb zircon age which agrees with the oldest K-Ar biotite age within analytical error. These volcanic and subvolcanic units should therefore be viewed as coeval eruptive/high-level intrusive equivalents of some of the late Triassic granitoid plutons. The result is significant as it represents the first documented case of late Triassic volcanism in the West Coast Province of Peninsular Malaysia.
With the exception of the 1250 + 300 Ma upper intersection age of the Tranum Granodiorite, all the upper intersection ages of the other reverse discordia chords fall in the range 1500 to 1700 Ma (Table 3). Reverse discordia may give unreliable estimates of old inherited zircon ages in cases where significant recent Pb loss has occurred. The presence of both internally and externally concordant zircon fractions, the consistency of the U-Pb zircon ages and the fact that no K-Ar or Rb-Sr mica age is older than the zircon crystallization ages yielded by the lower intersections of the reverse discordia are all indicative that negligible recent Pb loss has affected most of the inherited zircon patterns presented in this study. The reverse discordia, as such, may be conventionally regarded as resulting from mixing between magmatic zircon populations and old inherited zircons that have suffered variable but incomplete Pb loss during a melting event. Although the 1500 - 1700 Ma range may merely reflect the average age of source zircon populations of highly variable age, the regional consistency of the results is suggestive of incorporation of source zircons with a limited age range. It is consequently regarded that the relatively narrow 1500 - 1700 Ma upper intersection ages represent the ages of the most dominant zircon population in the source rocks of the West Coast Province granitoids. Whether such ages represent the age of the source rocks per se cannot be determined as the batholiths can be recognized as comprised of S-type granitoids derived from melting of metasedimentary source rocks. Such source rocks may contain detrital zircons much older than the stratigraphic ages of the host rocks. The results are however indicative of the sedimentary source rocks being derived in part by recycling of mid-Proterozoic granitoid/high-grade metamorphic basement rocks. In addition, the regional consistency of the inherited zircon ages
TABLE 3 U-Pb zircon regression and concordia intersection results for samples from the West Coast Province of Peninsular Malaysia.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Upper Intersection (Ma)</th>
<th>Lower Intersection (Ma)</th>
<th>Emplacement age indicated by U-Pb zircon system</th>
</tr>
</thead>
<tbody>
<tr>
<td>13</td>
<td>1560 ± 150</td>
<td>215 +5 -6</td>
<td>Lower intersection of 215±5 Ma.</td>
</tr>
<tr>
<td>14</td>
<td>1700 ± 600</td>
<td>209 ± 2</td>
<td>Lower intersection of 209±2 Ma.</td>
</tr>
<tr>
<td>29</td>
<td>1650 ± 400</td>
<td>215 +2 -4</td>
<td>Lower intersection of 215±2 Ma.</td>
</tr>
<tr>
<td>39</td>
<td>1610 ± 450</td>
<td>211 +5 -8</td>
<td>Lower intersection of 211±5 Ma.</td>
</tr>
<tr>
<td>38</td>
<td>-</td>
<td>-</td>
<td>Three concordant fractions give 198 ± 2 Ma.</td>
</tr>
<tr>
<td>41</td>
<td>-</td>
<td>-</td>
<td>Three concordant fractions give 206 ± 2 Ma.</td>
</tr>
<tr>
<td>45</td>
<td>1240 ± 300</td>
<td>208 ± 2</td>
<td>Lower intersection of 208±2 Ma.</td>
</tr>
<tr>
<td>78</td>
<td>1550 ± 350</td>
<td>219 +5 -9</td>
<td>Lower intersection of 219±5 Ma.</td>
</tr>
</tbody>
</table>
suggests that such basement rocks have a widespread distribution. More insights into the provenance of the West Coast Province granitoids are yielded by combined U-Pb zircon inheritance and Nd isotopic studies discussed in chapter six.

U-Pb ZIRCON DATING OF PLUTONS IN THE EAST COAST PROVINCE

Granitoid batholiths of the East Coast Province occur as narrow north-south trending masses aligned subparallel to the trace of the medial suture (Figs. 2, 18). The majority of the granitoids can also be recognized as I-types (Chappell and White, 1974) but scattered stocks of S-type granitoids can be recognized (Chapter five). A number of stratigraphic relationships may be broadly used to constrain the radiometric ages. The batholiths are intrusive into volcanic and sedimentary sequences regarded as Carboniferous to Triassic in age (Gobbett, 1973; Burton, 1973). Volcanic sequences interbedded with sediments that have been dated by fossils are regarded as coeval eruptive equivalents of the granitoid plutons. The Jurassic marked a period of uplift, deposition of flat-lying continental molasse sediments of post-orogenic character, and the end of the episodes of folding and granitoid intrusion. These field arguments suggest that the intrusive ages of the granitoid batholiths of the East Coast Province may be bracketed between the Carboniferous and Jurassic. Except for a few possibly reset ages younger than 190 Ma, the vast majority of the IGS K-Ar biotite and hornblende ages lie between 200 and 265 Ma and are thus in accord with regional stratigraphic constraints. Rb-Sr data of the IGS are more limited and based on regional whole rock isochrons. Two intrusive episodes at \( \sim 230 \) Ma and \( \sim 260 \) Ma have been suggested. Three plutons of greatest interest geochronologically have been selected for U-Pb zircon dating to check
Fig. 18 Distribution of granitoid batholiths in the northern portion of the East Coast Province (after Bignell and Snelling, 1977; Rajah et al., 1977). Locations of zircon samples studied are shown in relation to the two general age zones indicated by the work of Bignell and Snelling.
for consistency with the Rb-Sr and K-Ar results. Zircons separated from the three East Coast Province samples are similar to those of the West Coast Province in terms of their prismatic shapes, variable size and absence of obvious cores. Zircons from the two I-type plutons of the East Coast Province are, however, much more uniform in terms of colour. Zircons from these I-types are mainly of a transparent pale pink variety. Opaque brown grains commonly found in the high-U populations of the West Coast Province were not observed in the East Coast I-types.

Sample 132 (Fig.18) comes from an I-type granodiorite pluton yielding one of the oldest K-Ar biotite ages in the East Coast Province, 264 ± 7 Ma. This is, in the strict sense, a cooling age and the possibility of Ar loss or slow cooling makes this a minimum estimate of the age of emplacement. A U-Pb zircon date would be useful to test whether an older emplacement age (i.e. older than mid-Permian) may be possible. The four zircon fractions analysed display a reverse discordia pattern indicative of zircon inheritance (Fig.19). The fitted discordia chord (MSWD = 1.01) intersects concordia at 257±4 Ma and 790 ± 150 Ma. The indicated zircon crystallization age of 257±4 Ma lies within analytical error of the 264 ± 7 Ma K-Ar biotite age. It is possible, but not necessary, to suggest that minor Pb loss has shifted the zircon patterns to yield a younger lower intersection age. If such is the case, then the K-Ar biotite age would better reflect the emplacement age. In any case, both techniques imply a middle Permian (~260 Ma) emplacement age for the granodiorite pluton.
Fig. 19 U-Pb concordia diagram for zircon data of the East Coast Province plutons discussed in the text.
Sample 106 is from the S-type Kuantan Granite stock (Fig.18). The age of this stock is rather problematic. A K-Ar biotite age of \(217 \pm 8\) Ma was obtained by the IGS workers. Rb-Sr whole rock systematics indicate open-system behaviour or variable initial \(87\text{Sr}/^{86}\text{Sr}\) (Fig.20) but suggest that the stock belongs to the postulated 260 Ma intrusive episode. Results for three zircon fractions analysed are displayed on a concordia diagram in Fig.19. The reverse discordia is indicative of the presence of inherited zircon. A single zircon fraction plots on concordia within analytical error and yields an age of \(263 \pm 2\) Ma. Extrapolation of the discordia chord through the other points yields an upper intersection of \(1350 \pm 200\) Ma. The concordant age of \(263 \pm 2\) Ma is regarded as the best available estimate of the emplacement age of the Kuantan stock and is compatible with the much less reliable Rb-Sr whole rock result.

Sample 120 (Fig.18) was collected from a small, isolated pink-feldspar biotite adamellite pluton which is petrographically distinct from the adjacent biotite + hornblende granodiorite plutons of the Jengai and Maras-Jerong batholiths. Three Rb-Sr whole rock points yield an age of \(220 \pm 5\) Ma (Fig.21). This age is significantly younger than the 255 to 265 Ma ages of samples from the adjacent batholiths. This suggests that a number of small, late stocks found distributed along the coast within the older mid-Permian belt may constitute a distinctly younger minor intrusive episode. U-Pb zircon dating results are consistent with this suggestion. Four zircon fractions analysed from sample 120 show a normal discordance pattern attributable to post-emplacement Pb loss. One fraction falls on concordia to yield a zircon crystallization age of \(227 \pm 2\) Ma. The lower intersection is not significantly displaced from the origin. A recent Pb loss model is thus permissible although a more complicated
Fig. 20 Rb-Sr isochron plot showing a reference 252 Ma isochron fitted to five out of the seven whole rock analyses for the Kuantan Granite stock. Incorporates data from Bignell and Snelling (1977).

Fig. 21. Rb-Sr isochron plot for whole rock samples from the Kuala Dungun pink-feldspar adamellite. Incorporates data from Bignell and Snelling (1977).
scenario involving mass discriminatory loss is also possible. Such considerations do not, however, degrade the reliability of the crystallization age indicated by the concordant zircon fraction. The mid-Triassic age of 227 ± 2 Ma agrees well with the Rb-Sr isochron interpretation and confirms the presence of mid-Triassic plutons within the dominantly mid-Permian eastern zone of the East Coast Province.

The U-Pb zircon study of the three plutons from the East Coast Province highlights the reliability of the method for approximating emplacement ages. Results obtained support the main conclusions of the IGS K-Ar and Rb-Sr study. In all three cases documented in this section, U-Pb zircon ages agree well with either K-Ar mica or Rb-Sr whole rock isochron ages. In contrast to the plutons of the West Coast Province, this general concordancy of radiometric clocks is consistent with the high-level plutonic emplacement characteristics of the East Coast plutons (Hutchison, 1977).

Zircon inheritance patterns are documented for an I-type and an S-type pluton in the East Coast Province. Upper intersection ages obtained are 790 Ma and 1350 Ma respectively. These ages are evidence that these granitoids have either incorporated old crustal basement components of middle to late Proterozoic age or assimilated recycled zircons of this age present in the metasedimentary country rocks. Further evidence documenting the presence of underlying Proterozoic crust is discussed in chapter six.
ASPECTS OF ZIRCON INHERITANCE AND Zr SATURATION

The phenomenon of zircon inheritance in S-type granitoids has been well documented by various workers (this study, Pidgeon and Aftalion, 1978; Williams, 1977; Bickford et al., 1982 and others). Zircon populations from I-type granitoids, on the other hand, generally show little or no inheritance effects (e.g. Williams, 1977; Mattinson, 1978; Chen, 1982). The presence or absence of zircon inheritance effects correlating with the proposed twofold subdivision of orogenic granitoid batholiths is intriguing. Questions that arise include whether absence of inherited zircons in many I-types implies a non-zircon-bearing source and whether absence of inherited zircons in granitoids implies an absence of old underlying continental crust.

Watson (1979) discussed how Zr saturation effects vary with composition of felsic melts. As little as 100 ppm Zr is required to achieve saturation in peraluminous granitic compositions but several weight percent Zr is required for saturation in peralkaline felsic compositions. The experimental data also showed that Zr solubility possesses a strong temperature dependence. On the basis of their experimental results, Watson and Harrison suggested that non-peralkaline melts are likely to contain liquidus zircon crystals because the saturation level is so low for such compositions. It follows that old source zircons in such magma compositions would not be consumed during melting processes if Zr saturation is achieved. These broad conclusions are regarded as the basis for studying zircon behaviour in granitoids although complications may arise in cases where zircons occurring as inclusions in stable mineral phases are excluded from reacting with the melt and where source zircons are removed together with early fractionating phases in a magma undergoing
protracted fractionation. It is further assumed in subsequent discussions that if an old zircon survives a young melting episode, its old radiogenic Pb also survives.

Zr behaviour in Peninsular Malaysian granitoid suites

Fig.22A shows the behaviour of Zr on a Zr vs. SiO₂ variation diagram for the West Coast Province suites dated by the U-Pb zircon method. All the suites show consistently decreasing Zr with increasing silica indicating that zircon was an early liquidus fractionating phase which was continually removed. Such patterns are those expected of Zr saturation behaviour and are consistent with the general preservation of zircon inheritance in these suites. In this context solely concordant zircon populations for two plutons studied (samples 38,41) may have two plausible explanations - first, that protracted fractionation has removed essentially all of the old source zircons from the liquids forming the late members of a fractionation suite or secondly, that old source zircons were unstable in these two melt compositions. The similar chemistries of these two plutons with the other plutons studied and their high Zr contents compared to experimentally determined Zr saturation levels of Watson (1979) suggest that the former explanation is more probable.

The Zr behaviour of the East Coast Province suites is rather variable. Three broad patterns can be discerned. S-type suites, and I-type plutons like the Singapore Granodiorite (Fig.22B), show decreasing Zr with increasing SiO₂ and this pattern indicates early achievement of Zr saturation and probable preservation of old source zircons if such existed in the source regions. Sample 106, an S-type pluton belonging to this group does indeed possess zircons displaying zircon inheritance behaviour. A contrasting group of plutons
exemplified by the I-type granitoids of P.Ubin and the ~230 Ma
northern coastal stocks possesses unusually high Zr abundances in
spite of their high silica contents. This group plots significantly
above the main Zr vs. SiO$_2$ trend of the East Coast Province
granitoids. The P.Ubin rocks are associated with more mafic rocks
with lower Zr contents. This suggests that Zr increases with
increasing silica and Zr saturation was not achieved. Sample 120, a ~
230 Ma northern coastal stock, contains zircon populations that do not
show inheritance patterns. A third group, e.g. the Bt. Bari
Granodiorite (Fig.22B), displays an intermediate mode of behaviour.
Zircon inheritance has been documented for this pluton in a preceding
section. Zr displays an apparent maximum at ~70 to 72% SiO$_2$ and then
decreases. Such a pattern may be interpreted to indicate initially
undersaturated conditions and achievement of Zr saturation at
intermediate fractionation stages. The preservation of inherited
zircons may then be due to incomplete dissolution of old zircons,
sluggish dissolution rates, rapid achievement of saturated conditions
and isolation of zircons in minerals that prevented reaction with
melt.

Examination of Zr behaviour in I-type suites thus suggests that
zircon may be present as an early liquidus fractionating phase in some
suites but not in others. It may be predicted that even if source
zircons were originally present in the I-type source regions, these
would react with the melt and be consumed in a number of cases but not
in all cases.
Temperature: primary control for zircon inheritance preservation in orogenic granitoids?

Fig. 23 (after Watson and Harrison, in press) is a diagram showing Zr concentration levels required to achieve saturation for various melt compositions denoted by the melt basicity index \((\text{Na} + \text{K} + 2\text{Ca})/(\text{Al} \cdot \text{Si})\). Plotted on this diagram are some of the whole rock Zr concentrations and melt basicity indices of the Peninsular Malaysian granitoids, together with some southeast Australian I- and S-type granitoids with well-studied U-Pb zircon behaviour (Williams, 1977).

The plot shows that despite their rather different zircon inheritance features, there is no marked compositional differences that would explain why inheritance in I-types is rather uncommon. Zr concentrations in many I-types exceed the \(\sim 100\) ppm levels predicted as the general saturation limit for crustal melts by Watson (1979). A plausible explanation is that I-type sources, in general, lack zircon. This is not entirely satisfactory as some I-types (this study, Williams, 1977; Pidgeon and Aftalion, 1978) show inheritance, others show trace inheritance (e.g. Mattinson, 1978; Chen, 1982) and Nd isotopic work suggests that I-types with large negative \(\varepsilon_{\text{Nd}}\) values (e.g. chapter 6; DePaolo, 1981; McCulloch and Chappell, 1982) are crustal remelts. It is more probable that the 100 ppm Zr saturation limit of Watson is applicable for S-type granitoids but is too low in relation to I-type melts. Fig. 23 shows that for the temperature range 850 to 1000°C, which are reasonable liquidus temperature estimates for I-type magmas (e.g. Wyllie, 1977) of granodioritic-tonalitic compositions, from 300 to 1500 ppm Zr are required to achieve zircon saturation. A 100 ppm Zr saturation level implies liquidus temperatures of \(\sim 750^\circ\text{C}\), a typical estimate of metasedimentary crustal
Table 4: Some relevant chemical data used to describe Zr saturation behaviour in text.

<table>
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<tr>
<th>Sample</th>
<th>Zr (ppm)</th>
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<td>125</td>
<td>66.3</td>
<td>1.74</td>
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</tbody>
</table>
Table 23. Diagram showing variation of experimentally determined saturation curves vs. melt basicity index (after Watson and Harrison, 1983). Dashed curves are less reliably defined due to difficulties in obtaining low Zr concentrations by electron microprobe.
melting temperatures producing peraluminous S-type melts (e.g. Wyllie, 1977; Clemens and Wall, 1981). High Zr saturation levels for I-type melting conditions indicate that I-type crustal sources may contain old source zircons but these would generally be consumed during melting episodes.

U in zircons

A notable feature of the zircons studied in this work is their high U contents, in particular those of the S-type granitoids. U contents of the zircons range from 1000 to 5000 ppm (Table 1A-B). This range is significantly higher than the typical range of U contents (400 - 1000 ppm) in zircons from many granitoid terrains (e.g. Williams, 1977; Pidgeon and Aftalion, 1978; Mattinson, 1978; Chen, 1982). The high U contents recorded by the whole rock systems (Fig.24) thus reflect the intrinsically uraniferous nature of the granitoids and not postemplacement enrichment. In particular, the high U contents of the West Coast Province suites and the large volumes of these granitoids would suggest that this province constitutes an important regional U anomaly.

CONCLUSIONS

Precise estimates of the emplacement ages of felsic S-type granitoids like those of the West Coast Province batholiths of Peninsular Malaysia are rendered difficult by a number of factors. Variable Sr initial ratios, the possibility of preserving 'inherited isochrons', and open systems caused by hydrothermal activity commonly associated with felsic S-type granitoids all render the results obtained by the Rb-Sr whole rock isochron technique equivocal unless supported by other radiometric clocks. K-Ar and Rb-Sr mica ages may
Fig. 24. U contents of Peninsular Malaysian granitoids.
record younger postemplacement events such as resetting by younger plutons, slow cooling, hydrothermal resetting, or fault-related disturbances as are evident in the Peninsular Malaysian West Coast Province. Finally, Pb loss may interfere with zircon inheritance patterns so commonly documented for such S-type granitoids to yield complex patterns that seriously downgrade the potential of the U-Pb zircon clock. Nevertheless, the results of this study indicate that the U-Pb zircon clock yields by far, the most unambiguous and consistent estimates of the emplacement ages of the plutons studied. Patterns obtained show little or no Pb loss and indicate late Triassic (198 - 220 Ma) emplacement ages, generally \( \sim 10 - 20 \) Ma older than the oldest K-Ar and Rb-Sr mica ages. Results do not support previous suggestions that Palaeozoic granitoids are present in the two important regions of the West Coast Province that were studied. The view that Palaeozoic ages, less well documented in other parts of the West Coast Province, should be reinterpreted as late Triassic is adopted. The data also does not support Hamilton’s (1979) contention that the granitoids are largely of Jurassic age.

Limited U-Pb zircon work in the East Coast Province of Peninsular Malaysia essentially substantiates previous suggestions that the granitoid batholiths are of Permian to Triassic age.

Inherited zircon patterns have been documented for plutons from both provinces. Patterns obtained for the West Coast Province plutons are indicative that the source zircons incorporated during magma genesis are primarily dominated by 1500 - 1700 Ma populations. The two reverse discordia chords documented for the East Coast Province have \( \sim 800 \) Ma and \( \sim 1300 \) Ma upper intersections. Crystalline basement zircons, or recycled grains of such, older than 800Ma are therefore
indicated.

An examination of zircon inheritance behaviour in terms of Zr saturation characteristics of granitoids suggests that the preservation of inheritance patterns in S-type granitoids compared to their general paucity in I-types may be primarily the result of the higher temperatures attained during the production of I-type magmas. Other factors like differences in I-type and S-type compositional characteristics may be of minor importance.
CHAPTER FOUR

THE WEST COAST PROVINCE GRANITOIDS OF PENINSULAR MALAYSIA

INTRODUCTION

The West Coast Province granitoid batholiths of Peninsular Malaysia form an important segment of a long belt of batholiths, mainly of Mesozoic age, extending from Peninsular Malaysia northwards into northern Burma (Hutchison, 1977; Mitchell, 1977; Suensilpong, 1977; Pongsapich and Mahawat, 1977; Beckinsale, 1979). Mitchell (1977) reviewed the regional tectonic and geological positions of these batholiths and concluded that the Triassic batholiths in this belt were best explained to have formed in a continent collision setting. Hutchison (1978) and Beckinsale (1979) supported this model and Beckinsale suggested that most of the granitoids may be identified as S-type granitoids derived from melting of metasedimentary source rocks as defined by Chappell and White (1974). This chapter examines the mineralogical, chemical and isotopic characteristics of a number of suites from these postulated continent collision-related S-type batholiths in Peninsular Malaysia.

It is now becoming evident from recent studies and reviews of granitoid batholiths (Pitcher, 1979; Barker, 1981; Czamanske et al., 1981; Shaw and Flood, 1981) that S-type (as well as I-type) granitoids may display a wide range of mineralogical and chemical characteristics and that criteria adopted for identification of granitoid type in one terrain may not strictly hold true in another. In addition, suites with characteristics intermediate between the two end-member types have been recognized (Shaw and Flood, 1981). Such vagaries are to be expected when complexities involved in magma evolution and generation
processes are taken into consideration. For example, S-type suites of the Berridale and Kosciusko Batholiths of southeastern Australia differ in a number of ways from the isotopically primitive S-type suites of the New England Fold Belt of eastern Australia (Shaw and Flood, 1981) and continental collision-related Tertiary leucogranites of the Himalayas (LeFort, 1981; Dietrich and Gansser, 1981). Unlike the New England S-type suites, the Berridale-Kosciusko S-types and the Himalayan leucogranites are characterized by Nd and Sr isotopic signatures indicating derivation from old (>1000Ma) continental crust. The Himalayan leucogranites, however, differ from the Berridale-Kosciusko S-types with respect to absence of associated I-type granitoids, dominance of felsic high-SiO₂ compositions, absence of cordierite, presence of tourmaline + Mn-rich (?) garnet, presence of muscovite instead of biotite as the main mica and the abundance of associated aplites and pegmatites. The West Coast Province batholiths of Peninsular Malaysia share many of the characteristics of the Himalayan leucogranites but contain abundant biotite-bearing granodiorite-adamellite units and form large continuous batholiths that dwarf those of the Himalayan plutons. Aspects of the chemical and isotopic characteristics of the West Coast Province granitoids and their differences and similarities to other S-type granitoids are discussed in this study.

SAMPLING AND EXPERIMENTAL TECHNIQUES

Chemical and isotopic data for representative samples from eight granitoid suites and coeval volcanic/subvolcanic units in the West Coast Province are presented in this study. A number of polished thin-sections were selected after microscopic examination for limited electron microprobe studies to obtain mineral composition data. These
were obtained using an Si(Li) detector-equipped energy dispersive system at the A.N.U. Analytical procedures, interference corrections, precision and accuracy of this system are discussed by Reed and Ware (1975).

Much of the whole rock major and trace element data reported in this study were obtained by X-ray fluorescence spectrometry using Phillips 1220 and Phillips 1450 spectrometers. Major elements were determined on fused glass discs and trace elements on pressed powder pellets. Details of procedures, instrumental conditions and correction techniques are discussed by Norrish and Chappell (1977). FeO was determined by titration, Na2O by flame photometry and H2O and CO2 by gravimetric methods. REE, Cs, Hf and Ta abundances are also presented in this study and these elements were determined in duplicate by instrumental neutron activation techniques. Precision and accuracy of the INAA technique for the reported elements are estimated to be better than 10 percent.

Representative analyses of the S-type granitoids in the West Coast Province are shown in Table 1. Their sample locations are shown in Figs. 2 and 5. A complete list of all chemical analyses and description of sample locations may be found in the appendix.

GEOLOGICAL AND TECTONIC SETTING

The S-type granitoid batholiths of the West Coast Province are found to the west of a belt containing rocks of the ophiolite-melange association (Fig.1). This belt was termed the Bentong-Raub ophiolite line by Hutchison (1975, 1977). Mitchell (1977) interpreted this ophiolite line to mark a Triassic collision suture separating an eastern Malay Peninsula crustal block and a western Malay Peninsula
Table 1: Major and trace element abundances of representative samples of the West Coast Province granitoids.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>MAL7</th>
<th>MAL9</th>
<th>MAL13</th>
<th>MAL14</th>
<th>MAL20</th>
<th>MAL27</th>
<th>MAL29</th>
<th>MAL36</th>
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Trace elements

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Trace elements

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Fig. 1 Generalized distribution of granitoid batholiths in the West Coast Province of Peninsular Malaysia. Plutonic suites studied in this thesis are contained within the enclosed areas. These are shown in more detail in Fig. 2 and Fig. 5.
crustal block. He proposed that, in such a tectonic reconstruction, the granitoids of the West Coast Province were formed in a continent collision setting broadly analogous to that of the Tertiary Himalayan leucogranites (e.g., Dietrich and Gansser, 1981) found south of the Indus suture. This reconstruction is supported by the radiometric data discussed in chapter three and the mineralogical, chemical and isotopic data discussed in this chapter.

Hamilton (1979, p. 80) suggested that the granitoids of the West Coast Province were 'dominantly of Jurassic age and to a lesser extent, of Cretaceous and possibly also Triassic ages' and were primarily related to a Jurassic (?) - Cretaceous subduction system indicated by the ophiolite-melange belts of Burma to the west. This interpretation appears to be incorrect. U-Pb zircon results (chapter three) indicate that the West Coast Province batholiths are mainly and perhaps entirely of Triassic age. Data discussed in this chapter show that the characteristics of the West Coast Province granitoids stand in contrast to Mesozoic - Tertiary batholiths associated with subduction.

The S-type granitoid batholiths of the West Coast Province, occurring as north-south trending masses west of the suture, make up more than 50 percent of the total area of the province. Areal extent of the granitoids decreases markedly in a westward direction away from the suture. The huge Main Range Batholith, which is intrusive into rocks of the ophiolite-melange association, covers an area exceeding 15000 km². Granitoid complexes < 300 km² in area are found on the islands of Penang and Langkawi, 150 - 200 km west of the suture. Country rocks intruded by the granitoids are mainly of greenschist facies metamorphic grade. There is generally little increase in
metamorphic grade near plutonic contacts (Hutchison, 1978) and numerous faulted granitoid-metasediment contacts have been mapped by geologists of the Geological Survey of Malaysia. There are, however, very restricted localities where the metamorphism reached amphibolite facies grade adjacent to plutonic contacts e.g. G. Jerai Granite (Fig.1).

The sedimentary and metasedimentary formations intruded by the granitoids are of Cambrian to Triassic age and can be recognized as of continental-margin basin and shelf facies association. Volcanic rocks are rare and the very restricted volcanic/subvolcanic rocks displaying S-type characteristics that are coeval with the granitoids will be discussed in a short section at the end of this chapter.

BATHOLITH CONSTITUTION AND SUITE RECOGNITION

Two main batholith masses can be distinguished in the West Coast Province. These are the Main Range Batholith on the eastern flank and the adjacent Bintaang Batholith immediately to the west (Fig.1). Small intrusive centres are found further to the west. These are called the Kulim, Penang and Langkawi Complexes.* Each batholith/complex may consist of one or more granitoid suites. Such suites can be distinguished based on consistent mineralogical, chemical and isotopic differences. No detailed mapping has however been carried out to distinguish individual plutons comprising a suite. Presnall and Bateman (1973) and McCourt (1981) argued that, for I-type cordilleran batholiths like the Sierra Nevada and Peruvian Coastal Batholith, a suite results from high-level differentiation of a parent magma formed in a single fusion episode. This study will show that such an

* In this study, structurally continuous granitoid masses > 500 km² in area are termed batholiths whereas those that are < 500 km² are called complexes.
explanation can be plausibly extended to the S-type suites of the West Coast Province. Suites from two regions have been studied. These are:

1. the southern portion of the Main Range Batholith (Fig.2), and
2. the northwestern portion of the West Coast Province that includes granitoid suites from the Kulim Complex, Penang Complex and the southwestern lobe of the Bintang Batholith (Fig.5).

**Southern Main Range Batholith**

Four main intrusive suites separated by sedimentary screens can be distinguished in the southern portion of the Main Range Batholith. These are the Rembau Suite, Beranang Suite, Ulu Kali Suite and an undefined suite of which the Tranum Granodiorite is a constituent pluton (Fig.2). A major sedimentary screen and the subvolcanic Genting Sempah hypersthene microgranodiorite unit separate the Ulu Kali Suite from the Beranang Suite. A less prominent metasedimentary screen separates the Tranum Granodiorite from the adjacent Ulu Kali Suite. The recognized suites are typically ~10 km wide, ~50 km in length and are arranged subparallel to the main north-south batholith trend.

Rocks sampled from each suite are texturally quite variable. Rock types vary from biotite granodiorite to muscovite leucogranite. Porphyritic phases with large megacrysts of K-feldspar are common. In addition, sheared varieties, often displaying bluish quartz, are commonly observed in this portion of the batholith.
Fig. 2 Distribution of intrusive suites in the southern portion of the Main Range Batholith. Batholith outlines are based on data of Shu (1969), Gubbett (1972) and Gubbett and Tjia (1973).
Generalized characteristics of the suites studied are listed in Table 2. Fig. 3 A-D are plots of selected trace elements against SiO₂ and Fig. 4 A-B are Rb-Sr isochron diagrams that were found to be useful for discriminating between adjacent suites. The petrogenetic significance of the chemical and isotopic data will be discussed later. The important point to note here is that a variety of field, petrographic, chemical and isotopic data can be used to identify suites in the West Coast Province S-type batholiths with good consistency.

Radiometric dating of the suites has not yielded unequivocal results. Mica ages, whether by K-Ar or Rb-Sr, are invariably younger than ages determined by the U-Pb zircon method or the Rb-Sr whole rock isochron method. These young ages may plausibly be attributed to resetting by later thermal events or slow cooling. U-Pb zircon ages were considered to give the best estimates of the emplacement ages of the suites in chapter three. U-Pb zircon ages of the suites in this portion of the Main Range Batholith range from 198 Ma to 215 Ma.

**Bubu Suite (Bintang Batholith) and Kulim Complex (Fig. 5)**

A common feature of both these complexes is the presence of suites with the mineralogical assemblage low-Al biotite + sphene + actinolitic hornblende. These suites are associated with more typical high-Al biotite S-type plutons.

The southwestern lobe of the Bintang Batholith consists mainly of rocks of the Bubu Suite. The most common rock type is a mesocratic, coarse-grained porphyritic adamellite-granite (Yap, 1970). Low-Al biotite, sphene and actinolitic hornblende are present in the samples studied but occurrences of actinolitic hornblende seem to be irregular
Fig. 3. A-C Selected Harker variation diagrams illustrating the chemical fields for the adjacent Ulu Kali and Beranang Suites and the Tranum Granodiorite. Fig. 3D shows the distinctly lower $P_{2O_5}$ contents of the Rembau suite when compared with the adjacent Beranang Suite.
Fig. 4A  Rb-Sr isochron diagram for the Rembau and Beranang Suites. There is a distinct separation of the two suites - a higher Sr initial ratio of ~0.725 is indicated for the Beranang Suite compared with 0.710 for the Rembau Suite. A degree of open system behaviour is however, indicated for the Beranang samples which yield an unreliable age. A number of samples from the Beranang Suite with high $^{87}\text{Rb}/^{86}\text{Sr}$ ratios are not plotted. Data is from Bignell and Snelling (1977) except where marked by an asterisk which denotes a new determination.

Fig. 4B  Separation of samples from the Ulu Kali Suite and Tranum Granodiorite on a Rb-Sr isochron diagram. The initial ratio for the Ulu Kali Suite is 0.711, whereas for the Tranum Granodiorite it is 0.722. Data from Bignell and Snelling except for the Tranum sample with the highest $^{87}\text{Rb}/^{86}\text{Sr}$. 
<table>
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<th>Tranum</th>
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<td>Distinctive chemical features (Fig. 3)</td>
<td>Similar to Beranang suite but characteristic low FeO contents.</td>
<td>Low La, Th, Y. Negative trends for La, Th, Y vs SiO₂.</td>
<td>High La, Th, Y. Positive La, Y trends vs SiO₂ indicated.</td>
<td>Similar to Beranang</td>
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<td>Initial Sr ratio*</td>
<td>0.710 ± 0.002</td>
<td>Variable from 0.720 to 0.732. Possible open system behaviour for some samples.</td>
<td>0.711 ± 0.001**</td>
<td>0.722 ± 0.005</td>
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<td>Age determinations</td>
<td>221 ± 6 Ma Rb-Sr whole rock isochron. Oldest K-Ar biotite age is 184 ± 4 Ma.</td>
<td>211 ± 5 Ma and 215 ± 7 Ma U-Pb zircon ages. Large scatter for Rb-Sr whole rock method. High 87Sr/86Sr samples (&gt;97) yield model ages of 195 to 210 Ma for 0.72 initial ratio. Oldest K-Ar biotite age is 203 ± 5 Ma.</td>
<td>206 ± 2 Ma U-Pb zircon. 206 ± 2 Ma Rb-Sr whole rock isochron. Oldest K-Ar biotite age is 177 ± 4 Ma.</td>
<td>208 ± 2 Ma U-Pb zircon. 235 ± 30 Ma Rb-Sr whole rock isochron. Oldest K-Ar age is 165 ± 4 Ma.</td>
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* Bulk of Rb-Sr data obtained from Bignell and Snelling (1977). Interpretation modified slightly from Bignell and Snelling (1977) in light of some new data and suggested suite division. Sr initial ratios calculated from isochron intersections even when isochron ages do not agree with emplacement ages obtained by U-Pb zircon method. Preferred isochron ages are listed in Appendix 3.

** Sample 38 (Fig. 2) shows chemical affinities with the Ulu Kali Suite but yields an initial 87Sr/86Sr ratio of 0.748 if the Rb-Sr isotopic data are corrected using the 198 Ma U-Pb zircon age. It cannot be fully ascertained if this high initial ratio has resulted from open system behaviour or that sample 38 belongs to a separate suite that is chemically very similar to the Ulu Kali Suite.
(Yap, 1970). A separate medium-grained granite with high-Al biotite as the mafic mineral occurs along the southern margins of the lobe. Regression of the Rb-Sr whole rock data of Bignell and Snelling (1977) for the Bubu suite together with one new whole rock determination yields a model 4 age (appendix 3) of 198 ± 14 Ma and an initial ratio of 0.722 ± 0.003 (Fig.6). Three K-Ar biotite ages were reported by Bignell and Snelling (1977) for samples from the southern portion of the Bintang Batholith. A sample from the Bubu Suite yielded an age of 182 ± 5 Ma and the other two samples from adjacent suites yielded K-Ar biotite ages of 188 Ma and 205 Ma.

Courtier (1974) distinguished two separate granitoid intrusive masses in the Kulim Complex. In this study, these are referred to as the Penanti Suite and the Bongsu Suite. The Penanti Suite is located along the northwestern margin of the Kulim Complex and consists of low-Al biotite + sphene-bearing plutons, occasionally with rather distinctive pink K-feldspar phenocrysts. The Bongsu Suite includes the high-Al biotite plutons found south and east of the Penanti suite. Tourmaline, muscovite and garnet were reported by Courtier (1974) to be present in rocks of the Bongsu Suite. A model 1 age of 223 ± 23 Ma (appendix 3) is given by a best-fit line to eight Rb-Sr whole rock points for the Penanti Suite. The indicated initial $^{87}$Sr/$^{86}$Sr is 0.712 ± 0.003 (Fig.7). K-Ar biotite ages for the Penanti Suite reported by Bignell and Snelling are 198 ± 8 Ma and 203 ± 6 Ma. Two samples from the Bongsu Suite plot significantly above the array defined by samples of the Penanti Suite on a Rb-Sr isochron diagram (Fig.7). If a late Triassic age between 200 - 220 Ma is assumed, the indicated initial ratio for the Bongsu Suite is ~ 0.720. K-Ar mica ages range from 161 Ma to 192 Ma.
Fig. 5 Distribution of intrusive units studied in northwest Peninsular Malaysia. Diagram incorporates data from Gobbett (1972), Chung (1973) and Ng (1978).
Fig. 6 Rb-Sr isochron diagram for whole rock samples from the Bubu suite of the Bintang Batholith. Data from Bignell and Snelling (1977) except for one new sample denoted by asterisk.

Fig. 7 Rb-Sr isochron diagram for whole rock samples from the Penanti and Bongsu suites of the Kulim Complex. Data from Bignell and Snelling (1977) except where sample is marked by asterisk which denotes new data.
Penang Complex

Two suites, each ranging from adamellitic to leucogranitic compositions, are indicated by field mapping and geochemical data (Fig. 5, also chapter 3). These suites are called the Bt. Bendera and Sg. Ara Suites. The mineralogical assemblage (high-Al biotite + muscovite + tourmaline in addition to quartz and feldspar) and textural variations of both suites are similar to the suites of the southern Main Range Batholith discussed previously. Coarse-grained, porphyritic biotite (+ muscovite) adamellite-granite with K-feldspar megacrysts is the most common rock type. Pegmatites and aplites are ubiquitous late-stage phases associated with both suites.

Harker variation diagrams used to distinguish the two suites are shown in Fig. 8. The U-Pb zircon dating results discussed in chapter three indicate zircon crystallization ages between 209 Ma and 215 Ma but the results were not precise enough to distinguish different ages for each suite. K-Ar and Rb-Sr mica ages are all younger than U-Pb zircon ages and range from 68 Ma to 205 Ma. Rb-Sr whole rock data show considerable scatter on an isochron diagram (Fig. 9). If a late Triassic age of ~210 Ma is assumed, variable initial ratios between 0.710 and 0.720 are indicated for the Bt. Bendera Suite. The Sg. Ara Suite yields an excessively old Rb-Sr whole rock isochron age of 294 Ma which is inconsistent with U-Pb zircon and K-Ar mica ages. This result was suggested to represent an 'inherited isochron' age (chapter three). The initial ratio of 0.710 is, thus, of uncertain significance.
Fig. 8 A–E Major and trace element features of plutons of the Sg. Ara and Bt. Bendera Suites.
Fig. 9 Rb-Sr isochron diagram for whole rock samples from the Penang Complex. The Bt. Bendera and Gelugor plutons belong to the Bt. Bendera Suite; the Sg. Ara and Gertak Sanggul plutons belong to the Sg. Ara Suite. The 206 Ma 'reference isochron' joins the whole rock and K-feldspar points for one sample of the Bt. Bendera Suite. Regression of all samples of the Bt. Bendera Suite yields an 'age' of 177 Ma, a figure much younger than the 209 Ma U-Pb zircon age. The 294 Ma 'age' does not date emplacement of the Sg. Ara Suite. It represents an 'inherited isochron' age that is of uncertain significance (see text and also chapter 3). Data except where marked by asterisks are from Bignell and Snelling (1977).
Geochronology of the West Coast Province Suites

Many of the suites discussed previously have been dated using a number of radiometric methods that include U-Pb zircon, Rb-Sr whole rock, Rb-Sr biotite and K-Ar biotite or muscovite. Only in exceptional cases do two clocks yield ages that agree within analytical error. This leads to great difficulty in attempting to interpret which ages are the most reliable estimates of emplacement ages.

A number of generalizations can be made concerning the dating results (Bignell and Snelling, 1977; chapter three):

1. K-Ar and Rb-Sr mica ages range from 35 Ma to 210 Ma and are indicative of variable resetting by younger events. For results from the same pluton, these ages are always younger than U-Pb zircon ages and younger or equal to Rb-Sr whole rock isochron ages.

2. Rb-Sr whole rock isochron ages range from 198 Ma to 236 Ma (if the anomalously old 294 Ma age of the Sg. Ara Suite is excluded). Ages are often imprecise or even equivocal due to insufficient range in Rb/Sr ratios, variability of initial ratios and probable open system behaviour.

3. U-Pb zircon ages range from 198 Ma to 215 Ma. Post-emplacement open system behaviour is minimal. For the same suite, U-Pb zircon ages are older than K-Ar or Rb-Sr mica ages and agree with Rb-Sr isochron ages in the case of the Ulu Kali and Beranang Suites.
The present data base does not allow an unequivocal conclusion to be made with regard to emplacement ages of the various suites. Zircon crystallization ages yielded by U-Pb dating are the least ambiguous ages available. In cases like the Beranang and Sg. Ara Suites, emplacement ages can be bracketed between U-Pb zircon ages and the oldest K-Ar mica ages because, in a strict sense, these represent magmatic crystallization and cooling ages respectively. For the Beranang Suite, the emplacement age lies between 203 Ma and 215 Ma; for the Sg. Ara Suite, between 195 Ma and 215 Ma. Other suites cannot be treated in a similar fashion. For the Ulu Kali and Bt. Bendera suites, it is suggested that the U-Pb zircon ages between 198 Ma and 210 Ma be provisionally accepted as giving the best available estimates of emplacement ages. In the case of the Rembau, Bubu, Penanti and Bongsu Suites for which U-Pb zircon ages are not available, Rb-Sr whole rock isochron ages or K-Ar mica ages that lie between 200 Ma and 220 Ma probably represent the best available estimates of emplacement ages.

GENERAL MAJOR ELEMENT CHEMISTRY

The West Coast Province suites studied display a number of chemical features that may be regarded as typical of S-type granitoids. On a conventional AFM diagram (Fig.10), the samples plot in a rather limited field near the Na₂O + K₂O corner. On a modified ACF diagram (White and Chappell, 1977), the samples plot mainly in the area between the plagioclase - biotite - cordierite tie lines but some highly peraluminous compositions plot above the plagioclase-cordierite tie line (Fig.11). Molecular Al₂O₃/(Na₂O + K₂O + CaO) for the majority of the samples range from 1.03 to 1.23 (Fig.12). Notable exceptions are a handful of sphene + actinolitic
Fig. 10 Clustering of the West Coast Province granitoids near the Na$_2$O + K$_2$O corner on an AFM diagram.
Fig. 11 West Coast Province granitoids plotted on a modified ACF diagram (White and Chappell, 1977). Peraluminous granitoids plot above the plagioclase - biotite tie line. A=atomic (Al-Na-K); C=atomic (Ca-P); F=atomic (Fe+Mg). Symbols as for Fig. 10.
hornblende-bearing samples that possess ratios between 0.95 and 1.02. $\text{SiO}_2$ contents of the analysed samples range from 66% to 77% but most are in the 70-77% $\text{SiO}_2$ range.

Aspects of the major element chemistry of the West Coast Province suites are best discussed by comparison with other S-type batholiths so that some perspective of the variations in chemical characteristics of S-type granitoids can be realised. Fig.13A-G compares some of the major element trends of the West Coast Province suites with the S-type suites of the Kosciusko Batholith (Hine et al., 1978) and the Tertiary leucogranites of the Bhutan Himalayan (Dietrich and Gansser, 1981). The choice of these two terrains is based on their contrasting settings and mineralogical characteristics. The former is an example of a Palaeozoic continent-margin batholith containing associated I-type plutons. The ferromagnesian assemblage of these S-type plutons is biotite + muscovite + cordierite. The Himalayan leucogranites are thought to be generated during continent collision; they consist exclusively of S-type granitoids and the ferromagnesian assemblage is muscovite + biotite + tourmaline ± (Mn-rich?) garnet. For comparable $\text{SiO}_2$ contents, the Himalayan leucogranites show the lowest CaO, FeO, MgO and the highest Na$_2$O and Al$_2$O$_3$ contents whereas the Kosciusko rocks show the inverse features. The latter also show the lowest K$_2$O contents. The high Na$_2$O contents of the West Coast Province and Himalayan rocks displayed on an Na$_2$O vs. K$_2$O plot (Fig.13 G) is a notable feature indicating high-Na sources for these granitoids. All three S-type provinces are characterized by low Fe$^{3+}$/total Fe values that reflect the typically reduced nature of the source rocks of S-type granitoids. The Fe$^{3+}$/total Fe values for the West Coast Province suites range from 0.2 to 0.6 (Fig.14) but a few highly evolved leucogranites and microgranites show higher values.
Fig. 12  A = Plot of A/CNK (=mol. Al$_2$O$_3$/[Na$_2$O + K$_2$O + CaO]) vs. SiO$_2$ showing increasingly peraluminous nature of highly differentiated compositions.  B = Comparison of A/CNK values of the West Coast Province granitoids with those of the Kosciusko Batholith on a histogram. Symbols as for Fig. 10.
Fig. 13 A-F Major element oxide trends of the West Coast Province granitoids plotted on Harker variation diagrams. The fields for the Tertiary leucogranites of the Bhutan Himalayan (Dietrich and Gansser, 1981) and the Kosciusko Batholith of the Lachlan Fold Belt (Hine et al., 1978) are reproduced for comparison. Fig. 13G compares the Na$_2$O - K$_2$O fields of these granitoid terrains. Symbols as for Fig. 10.
Fig. 13  See previous page
Fig. 13G Na$_2$O vs. K$_2$O plot of granitoid samples from the West Coast Province. Also shown are the generalized fields for S-type granitoids of the Kosciusko Batholith and the Bhutan Himalayan region (see previous Figures).
Fig. 14 Variation of Fe$^{3+}$/total Fe (atomic proportions) with SiO$_2$ for the granitoids of the West Coast Province. Some of the high SiO$_2$ granities (>74% SiO$_2$) showing high Fe$^{3+}$/total Fe values display characteristics indicating reaction with a fluid phase. Highly oxidized, magnetite-bearing I-type granitoids typically show Fe$^{3+}$/total Fe values > 0.6.
between 0.6 to 0.8.

All three S-type provinces discussed show fundamental S-type features like peraluminous compositions, high $\text{K}_2\text{O}/\text{Na}_2\text{O}$, low $\text{Fe}^{3+}/\text{total Fe}$ and a restricted composition range dominated by high-$\text{SiO}_2$ granitoids. There are, however, significant inter-province chemical differences which may reflect differences in source characteristics as well as effects of magma generation and differentiation processes. The high $\text{K}_2\text{O}$ contents and high $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratios of the West Coast province granitoids compared with the Himalayan and Kosciusko granitoids probably reflect differences in source characteristics - a plausible explanation is a greater proportion of high K/Na clay minerals like illite and hydromuscovite in the sedimentary sources.

The West Coast Province granitoids are not as strongly peraluminous as the Kosciusko Batholith S-types - this is evident from the histogram plot of mol. $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$ (Fig. 12). The highly peraluminous nature of the Kosciusko Batholith S-types is consistent with the common occurrence of cordierite in these rocks. If the cordierite largely represents restite (White and Chappell, 1977), then the high mol. $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$ values of the Kosciusko Batholith S-types may be explained as the result of the presence of highly peraluminous restite phases. Actual liquid (or minimum melt of White and Chappell, 1977) is not highly peraluminous. For example, a typical S-type minimum melt of White and Chappell (1977) has a mol. $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$ value of 1.07. For the case of S-type liquids that crystallize early cordierite, garnet or even sillimanite, melts must be initially saturated in $\text{Al}_2\text{SiO}_5$ (Clemens and Wall, 1981). Such melts are derived by partial melting
of highly peraluminous source rocks. The moderately peraluminous nature of the bulk of the West Coast Province granitoids and the absence of cordierite, Fe-Mg garnet or sillimanite is most consistent with derivation from metapelitic source rocks undersaturated with respect to Al$_2$SiO$_5$.

An additional mechanism is however required to explain a number of anomalously peraluminous felsic compositions. Although most of the high-SiO$_2$ granites of the West Coast Province possess mol. Al$_2$O$_3$/(Na$_2$O + K$_2$O + CaO) values between 1.03 and 1.15, a small number of highly evolved leucogranites and microgranites have high values exceeding 1.15. These samples display other anomalous chemical characteristics suggesting that they have evolved by reaction with a fluid phase (see later section on high-SiO$_2$ granites) at some time and may have lost Na$_2$O and K$_2$O relative to Al$_2$O$_3$. Loss of alkalis would result in high mol. Al$_2$O$_3$/(Na$_2$O + K$_2$O + CaO) values and represents another plausible mechanism to generate highly peraluminous compositions. Its relevance is however restricted to the highly evolved high SiO$_2$ granites in the West Coast Province and its importance as a general mechanism to produce peraluminous compositions is regarded as minor.

MINERAL CHEMISTRY

Feldspars

Plagioclase displays a variety of habits. It may occur as discrete phenocrysts or as glomeroporphyritic aggregates showing resorbed outlines in the mafic members of a suite. Such grains are invariably zoned and contain calcic cores. The morphology and compositions of these grains suggest that plagioclase was an early
liquidus phase having a reaction relationship with evolving melts. Complexly zoned plagioclases with calcic, altered and inclusion charged cores considered by White and Chappell (1977) as diagnostic of restite plagioclase is not common and seems to be restricted to the mafic members of the suites. In contrast to these habits that indicate early crystallization of plagioclase, late sodic plagioclase may be found together with quartz and K-feldspar making up an even-grained groundmass in felsic units. Plagioclase also occurs as inclusions in K-feldspar megacrysts and as exsolution albite patches in K-feldspar.

Fig. 15 shows a generalized plot of plagioclase compositions against whole rock SiO$_2$ contents. Vertical bars represent the range in composition found in zoned crystals estimated from core and rim determinations. Although there is a clear trend of decreasing An content of plagioclase with increasing whole rock SiO$_2$, it is evident that a number of high-SiO$_2$ units contain entrained grains with relatively calcic cores.

Plagioclase compositions range from An$_{52}$ to An$_0$. Zoning in a single thin-section may exceed 20 mol. % anorthite. The most common plagioclase is oligoclase. The plagioclase trend shown in Fig.15 appear to flatten out at around An$_{45}$ to An$_{50}$ in the more mafic units. This feature, together with the An$_{50}$ plagioclases of volcanic-hosted granular xenoliths having plagioclase + hypersthene + cordierite + biotite + spinel assemblages that are interpreted as representing restite xenoliths, suggest that An$_{45-50}$ may be considered as a good estimate of restite plagioclase composition for the West Coast Province suites. Plagioclase of albite composition is found in leucogranites and felsic microgranites, as exsolution patches
Fig. 15 Plot of range of plagioclase compositions vs. whole rock SiO$_2$ content for selected samples of the West Coast Province granitoids. Vertical bars represent extremes of An contents in cores and rims of zoned crystals. Sample 79 is a subvolcanic hypersthene microgranodiorite showing 'S-type' chemical features.
in K-feldspar and as occasional thin rims around more calcic plagioclase grains. The latter is interpreted as a late-stage alteration feature.

K-feldspars occur as megacrysts in coarse-grained units, as early phenocrysts in a number of microgranites, as equigranular groundmass grains and as late interstitial patches. Exsolution textures of varying scales are ubiquitous. Occasional rapakivi-type oligoclase mantling of K-feldspar phenocrysts is present in a number of plutons but is not a unique characteristic of any single suite. K-feldspar megacrysts invariably contain inclusions of plagioclase, biotite and quartz. Compositions of analysed K-feldspars range from Or\textsubscript{90} to Or\textsubscript{100} and x-ray diffraction studies (Hutchison, 1977) indicate that the K-feldspars are invariably maximum microcline.

Biotite

Biotite may occur as discrete plates, as ragged shreds in mafic clots and as small flakes associated with granoblastic aggregates of quartz and plagioclase. The pleochroism scheme is typically pale-brown to reddish-brown. Pods of epidote and chlorite often occur along the cleavages of biotite. Occasionally, secondary granules of sphene are found rimming biotites.

All biotites analysed (Table 3) are more aluminous than the phlogopite–annite series (Fig.16A). Atomic Al\textsuperscript{4} varies from 2.3 to 2.7 and atomic Fe/(Fe+Mg) varies from 0.5 to 0.85. Iron-rich compositions trend towards the Al-rich siderophyllite corner (Fig.6A). The highly aluminous nature of the West Coast Province biotites is however not well displayed on an Al\textsuperscript{4} vs. Fe/(Fe+Mg) plot - this is apparent from the overlap between the field for the West Coast
Table 3 Representative mineral composition data obtained by electron microprobe.

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'ilm' (%) | 99          | 100          | 93           | 95           | 100           |
Fig. 16A  Plot of $\text{Al}_4^+$ vs. $\text{Fe}/(\text{Fe} + \text{Mg})$ contents of biotites from the West Coast Province granitoids. Field of I-type granitoids from the East Coast Province of Peninsular Malaysia from chapter 5.
Fig. 16B Plot of $\text{Al}^6$ (atomic Al remaining after subtracting amount required to balance $(\text{Si}^4 + \text{Al}^4) = 8$) vs Fe/(Fe + Mg) of biotites from the West Coast Province granitoids. Symbols as for Fig. 16A.
Province biotites and the field for the biotites from the East Coast Province I-type granitoids (Chapter five). A plot of $\text{Al}^6$ vs. $\text{Fe}/(\text{Fe} + \text{Mg})$ (Fig.16B) shows the clear separation of the high-$\text{Al}$ biotite compositions of the West Coast Province granitoids from the low-$\text{Al}$ biotites of the East Coast Province I-type granitoids. An exception is the field of the biotites from some sphene + actinolitic hornblende-bearing plutons of the Kulim Complex and the Bubu Suite. Biotites from these plutons show low to intermediate $\text{Al}^6$ contents. The biotite data and the presence of Ca-bearing phases like sphene and actinolitic hornblende indicate that the source rocks for these plutons may contain more calcareous and less pelitic components compared with the other West Coast Province plutons.

$\text{Mg}/(\text{Mg}+\text{Fe})$ ratios of biotites decrease with increasing whole rock $\text{SiO}_2$ (Fig.17). Biotites from the sphene + actinolitic hornblende-bearing plutons are more Mg-rich for comparable whole rock $\text{SiO}_2$ contents than other West Coast Province plutons. MnO contents (0.1 - 0.6 wt. %) are rather typical of biotites from felsic granitoids (Miller and Stoddard, 1981). MnO displays an overall negative correlation with $\text{Mg}/(\text{Mg}+\text{Fe})$ (Fig.18). If biotite MnO contents are reflecting those of the coexisting melts, this pattern could be interpreted to indicate that the melts are enriched in MnO with increasing differentiation. There is however little correlation for intra-suite samples between MnO and $\text{Mg}/(\text{Mg}+\text{Fe})$ ratios and it is more probable that decreasing amounts of coexisting manganiferous ilmenite is the main reason for the biotite MnO vs. $\text{Mg}/(\text{Mg}+\text{Fe})$ overall trend. The implications of these mechanisms for the existence of Mn-rich garnet in some leucogranite compositions is considered in a later section.
Fig. 17 Plot of Mg/(Mg + Fe) ratios of biotites against whole rock SiO₂ contents showing the Fe enrichment in biotites with increasing differentiation. Symbols as for Fig. 16A.

Fig. 18 Plot of wt. % MnO vs. Mg/(Mg + Fe) of biotites for the West Coast Province granitoids. Although there is an overall trend of increasing MnO content with decreasing Mg/(Mg + Fe) ratios, there is little correlation of these parameters for samples from the same suites. Symbols as for Fig. 16A.
Ohmoto and Kerrick (1977) discussed devolatilization equilibria in graphitic systems and their effects on granite minimum melting equilibria. They showed that stability curves of biotites ranging in composition from pure annite to phlogopite all lie below their 'maximum graphite' stability curve ($\log f_0^2 \sim -18$ at 700°C, 2 kb) if biotite coexisted with a C-O-H-S fluid system in equilibrium with pyrite, pyrrhotite and graphite. The equilibria are considered to be relevant for reduced S-type magmas lacking magnetite and containing pyrite and pyrrhotite like the West Coast Province granitoids. $f_0^2$ conditions of the West Coast Province granitoids are consequently believed to lie below the 'graphite maximum' curve of Ohmoto and Kerrick (1977).

Biotite is an early crystallizing phase in the West Coast Province granitoids. Textural relationships in the mafic members of the suites indicate that biotite crystallized before quartz and K-feldspar. Occasional inclusions of biotite in zoned plagioclase were observed in thin section. Crystallization sequences indicated by such textural relationships yield valuable information pertaining to melt $H_2O$ contents. The 1-2 kb results of Clemens and Wall (1981) for an S-type granodiorite indicate that water contents of the melt phase must exceed 2.5 – 3.5 wt. % for biotite to form before quartz and K-feldspar, and exceed 4 wt. % for biotite to form earlier than plagioclase. If the absence of orthopyroxene in the granitoids of the West Coast Province is interpreted as indicative of low-temperature magmas (< 800°C), the inferred crystallization sequence suggests water contents exceeding 5 wt. % in the melt phase. High water contents even in the early unfractonated melts of the West Coast Province suites are indicated. The dangers, however, of extending the experimental results too literally to complex geological situations
and differing granitoid compositions are well appreciated. Nevertheless, even if actual values of melt water contents may not be accurately estimated, the qualitative conclusion that the West Coast Province magmas had high initial water contents is not compromised.

Muscovite

Muscovite may occur as texturally 'primary' flakes in early granodiorite plutons, as large, interstitial plates in leucogranites and felsic microgranites and as dispersed secondary patches associated with alteration of feldspars. The relatively coarse, euhedral flakes occasionally containing apatite and opaque inclusions and not associated with feldspars are regarded to be primary. Such flakes are present in minor amounts (< 5 vol. %) in the granodiorites and adamellites of the Sg. Ara, Bt. Bendera and Beranang suites. In most cases, however, the origin of the muscovite observed in thin section is difficult to decipher when it is not obviously associated with sericitic alteration of feldspar. It is probable that in most samples, the muscovite present comprises of complex assemblages of magmatic and high-temperature subsolidus populations. The analyses discussed in this section excludes muscovite grains associated with feldspar alteration but includes grains of primary magmatic origin as well as coarse interstitial muscovite plates of uncertain genesis.

Fig. 19 A–C shows some of the important chemical characteristics of the analysed muscovites on plots suggested by Clarke (1981), Miller et al., (1981) and Anderson and Rowley (1981). Fig. 19 A shows that all of the muscovites analysed contain significant amounts of celadonite \( (K_2Al_2(Fe,Mg)_2Si_8O_{20}(OH)_4) \) and ferrimuscovite \( (K_2Fe_4(Si,Al)_2O_{20}(OH)_4) \) components. The amounts of (celadonite + ferrimuscovite) range from 10 to 30%. Fig. 19 B shows
Fig. 19A Atomic Si-Al-other diagram used by Miller et al. (1981) to estimate proportions of celadonite, ferrimuscovite and muscovite components. The West Coast Province 'muscovites' contain 70 - 90% of the ideal muscovite end member.
Fig. 19B Plot of \( \text{Na}/(\text{Na} + \text{K} + \text{Ca}) \) vs. \( \text{Fe} + \text{Mg}/(\text{total octahedral sites}) \) of analysed muscovites. Diagram illustrates extent of deviation from ideal muscovite compositions (where both ratios = 0) and proportions of paragonite and (ferrimuscovite + celadonite) components in the analysed muscovites. Plot is that of Clarke (1981).

Fig. 19C Plot of \( \text{Al}^6 \) vs. wt. \% \( \text{TiO}_2 \) of muscovites from the West Coast Province granitoids. As exemplified here by \( \text{Al}^6 \), there are no chemical parameters that consistently discriminate between high-\( \text{TiO}_2 \) (primary?) muscovites with \( \text{TiO}_2 > 0.4 \) wt. \% and low-\( \text{TiO}_2 \) secondary muscovites.
that the amount of the paragonite \((\text{Na}_2\text{Al}_4(\text{Si}_6\text{Al}_2\text{O}_{20})(\text{OH})_4)\)
component vary from 0 to 12%. Fig. 19 C, a plot of \% TiO_2 against
Al^6, may be a useful diagram to use for deciphering the origin of
the muscovites.

Miller et al. (1981) and Anderson and Rowley (1981) suggested
that primary muscovite is generally characterized by high TiO_2
contents, typically exceeding 0.5%. The low TiO_2 contents of most
of the muscovite compositions (mostly <0.3 wt. % TiO_2) shown in
Fig. 19 C are suggestive that muscovite is mainly of secondary origin.
A number of grains thought to be primary, based on textural grounds,
have TiO_2 contents of 0.7 to 1.2 wt. %. These primary grains show
the highest paragonite (\(~10\%) and lowest celadonite + ferrimuscovite
(\(~10-20\%\)) contents. Low-TiO_2 muscovites from highly evolved
leucogranites and microgranites, however, also display similar high
paragonite and low celadonite + ferrimuscovite contents. Although it
is probable that the TiO_2 contents of the muscovites are more
reliable indicators of primary vs. secondary origin, such variable and
decoupled chemical behaviour indicate that unequivocal compositional
discrimination between primary and secondary muscovite is difficult.

The suggested presence of primary muscovite in some of the West
Coast Province suites does not imply a high pressure (> 3-4 kb)
magmatic origin. The presence of celadonite and ferrimuscovite
components in a muscovite may increase the stability field of
muscovite to lower pressures around 2.5 kb or possibly even less

Muscovite is restricted to plutons containing high-Al biotite and
is absent from the sphene + actinolitic hornblende - bearing plutons
of the Kulim Complex and Bubu Suite (only minor sericitic alteration
is observed). Muscovite, therefore, is a reliable indicator of
peraluminous magma compositions in the West Coast Province,
irrespective of whether the muscovite is primary or of subsolidus
origin.

The presence of primary muscovite can be used to estimate water
contents of the melt phase when muscovite was formed. For reasonable
temperatures around 700 - 750°C and pressures around 3 - 5 kb, the
experimental results of Huang and Wyllie (1981) indicate melt water
contents of 6 to 10 wt. %. These estimates are qualitatively
consistent with melt water contents estimated based on biotite
crystallization sequence arguments discussed in the previous section.

Tourmaline

Tourmaline commonly occur as interstitial masses and skeletal
networks which are most apparent in leucogranitic compositions.
Grains were also observed replacing biotite and in a single case, the
crystals formed a radiating pattern. In all cases, the very
late-stage nature of the tourmaline is undeniable and all are probably
of metasomatic origin. Electron microprobe studies show that the
tourmalines belong to the schorl (Fe-rich) - dravite (Mg-rich)
solid-solution series. FeO + MnO/(FeO + MnO + MgO) values range from
0.73 to 0.92. The tourmaline compositions are compared with
tourmalines from two similar granitoid terrains on an MgO - TiO₂ -
(FeO + MnO) diagram in Fig.20 (after Manning, 1982). In common with
the tourmalines from the southwestern English and southern Thai
granitoids, the West Coast Province tourmalines show a trend from
relatively dravite-rich compositions in less differentiated rocks to
schorl-rich compositions in highly differentiated compositions
exemplified by the high-SiO₂ leucogranites.
Fig. 20 Compositions of tourmalines from the West Coast Province granitoids plotted on (FeO + MnO) - MgO - TiO₂ triangular diagrams and compared with Fe-rich tourmaline compositions of other 'tin granite' areas. Diagrams for the Cornish suite and the Hub Kapong Batholith of southern Thailand are from Manning (1982).
Garnet

Garnet was observed in only two samples. Both these samples are from the Beranang Suite and are high-SiO₂ leucogranites that display a number of anomalous chemical features like unusually high mol. Al₂O₃/(Na₂O + K₂O + CaO), high Fe³⁺/total Fe (>0.5) and high Mn. Garnet makes up less than 5 modal percent in thin-section for both samples. The observed garnets are texturally quite variable, even in a single thin section. They may occur as small (<0.5 mm), inclusion-free euhedra in quartz and albite, as larger (>1 mm) grains closely associated with muscovite-quartz-albite aggregates and as late interstitial grains containing quartz inclusions and also associated with muscovite, quartz and albite. In both garnet-bearing samples, tourmaline is present and biotite is absent.

Microprobe analyses show that the garnets are high-Mn members of the spessartine (Mn₃Al₂Si₃O₁₂) - almandine (Fe₃Al₂Si₃O₁₂) series (Fig.21). Zoning is pronounced and grains are zoned to high-Mn rims (up to 60 mol. % spessartine). The amounts of grossular, andradite and pyrope components are uniformly low and total less than 5 mol. % combined. The high spessartine contents of the garnets differ very significantly from the low spessartine (~10 mol. %) garnets of the South Mountain Batholith (Allan and Clarke, 1981), Strathbogie Batholith (Phillips et al., 1981) and liquidus garnet compositions of Green (1977) and Clemens and Wall (1981).

The garnets are obviously not of restite origin or country rock xenocrystal origin. Their compositional features, mode of occurrence and restriction to leucogranitic plutons rule out these possibilities. Three other mechanisms may be considered. (1) crystallization from a
Fig. 21 Plot of analysed garnet compositions (solid circles) on an atomic Mn-Fe-Mg triangular plot. The high Mn compositions of the garnets are compared to more typical Fe-rich garnets from other granitoid suites studied. Data sources = north Portugal pegmatites and aplites, Neiva (1975); Old Woman-Piute Range, Miller and Stoddard (1981); South Mountain Batholith, Allan and Clarke (1981); Strathbogie, Phillips et al. (1981); Werner Batholith of Antarctic Peninsula, Vennum and Meyer (1979); Ruby Mountains of Nevada, Kistler et al. (1981).
highly differentiated melt enriched in Mn with high Mn/(Mn + Fe + Mg) ratios (Clarke, 1981 and references cited); (2) reaction between crystalline phases and evolving melt of the type, liquid + biotite = garnet + muscovite or liquid + Al$_2$SiO$_5$ + biotite = garnet (Clarke, 1981 and references cited); and (3) replacement of feldspar by late-stage fluids described by reactions like:

$$2\text{NaAlSi}_3\text{O}_8 + 3(\text{Mn,Fe})^{2+} + 2\text{H}_2\text{O} \rightarrow (\text{Mn,Fe})_3\text{Al}_2(\text{SiO}_4)_3 + 2\text{Na}^+ + 3\text{SiO}_2 + 4\text{H}^+$$

albite fluid spessartine

(Plimer, 1974). The H$^+$ produced by such reactions leads to breakdown of K-feldspar to produce muscovite and quartz.

The complexities of transition processes between silicate melt + crystals, silicate melt + crystals + vapour, crystals + vapour, silicate melt + vapour may blur distinctions between mechanisms (1) and (3). A number of lines of evidence suggest that mechanism (3) may be the most plausible explanation for the paragenesis of the garnets. These include the unusual chemistry of the garnet-bearing rocks (reaction with an oxidized, volatile-rich fluid phase would explain high rare metal contents including Mn, Ga, Nb, high Fe$^{3+}$/total Fe and loss of alkalis would result in highly peraluminous compositions), presence of tourmaline, high-spessartine contents of the garnets and the association of garnet-muscovite-albite-quartz. Late-stage replacement, however, may not be a general mechanism to explain the origin of all spessartine-rich garnets in felsic granites. Miller and Stoddard (1981) have suggested that mechanism (2) best explains the paragenesis of the spessartine-rich garnets of the Old Woman – Piute Range peraluminous plutons of California.
Amphibole

Minor amphibole, occurring as occasional prisms, subhedral sieve-textured grains and shreds associated with mafic clots, is restricted to suites within the Bintang Batholith and adjacent satellitic plutons (this study; Yap, 1970; Santokh Singh and Yong, 1982). Analyses of the amphiboles coexisting with sphene in plutons of the Bubu Suite show that these are actinolitic hornblendes (Fig. 22). Atomic Mg/(Mg + Fe) range from 0.5 to 0.6.

The presence of actinolitic hornblende in the Bubu Suite plutons is interpreted as resulting from the more calcareous nature of the source rocks of this suite compared with the more typical S-type suites of the Main Range Batholith. Other occurrences of amphibole in suites recognized as S-type have been documented. Shaw and Flood (1981) reported the presence of actinolitic hornblende and cummingtonite in a number of S-type plutons of the New England Batholith in eastern Australia.

Opaque mineralogy

A general feature of the suites studied is their low opaque mineral content. Opaque minerals may range up to several modal percent in mafic granodiorites and adamellites but are much less abundant in more felsic compositions. The primary opaque minerals are often found as anhedral, discrete masses associated with mafic clots or as inclusions in biotite and sphene. These are quite distinct from occasional fine-grained aggregates associated with altered biotite. Ilmenite, pyrite and pyrrhotite are the opaque minerals present. Of these, ilmenite is the most common opaque phase. No exsolution features were discernible in the ilmenites in polished thin-section.
Fig. 22 Compositions of amphiboles in granitoids of the Bubu Suite (Bintang Batholith) plotted on Leake's (1978) amphibole nomenclature diagram. Atomic (Na + K)\(_A\) is $< 0.5$ and atomic Ti $< 0.5$ for analysed amphiboles calculated based on 23 oxygens following Leake (1978).

Fig. 23 Plot of atomic Mn in ilmenite vs. atomic Mn in coexisting biotite for the West Coast Province granitoids.
No magnetite was found in the polished thin-sections studied.

All ilmenites analysed contain significant amounts of MnO (from 2 to 8 wt. %). Calculated mol. % ilm [FeTiO_3/(FeTiO_3 + Fe_2O_3)] for sixteen samples containing ilmenite range from ilm_98 to ilm_100. In contrast, two samples from the sphene-bearing Penanti Suite (Kulim Complex) and one sample from a sphene-bearing pluton of the Bubu Suite have ilmenites with a composition range of ilm_92 to ilm_95. The ilmenite data need to be interpreted with caution although the implication that the sphene-bearing plutons were derived from relatively more oxidized magmas because of the presence of sphene and higher Fe_2O_3 contents of the ilmenites is compelling. Other explanations are possible and include: (1) plutons of the Penanti and Bubu Suites have undergone late-stage oxidation processes and these were recorded by the ilmenites, and (2) the ilm_98 - ilm_100 compositions obtained from sphene-absent plutons record reequilibration to reduced conditions during slow cooling. Ilm_99 compositions from subvolcanic rocks and fine to medium-grained phases that have presumably been relatively rapidly cooled are, however, suggestive that the ilmenites may be recording magmatic conditions.

Ewart (1981) reported near stoichiometric ilmenites without coexisting spinel in some Tertiary felsic volcanics of Queensland, Australia and suggested that this feature indicates equilibration under reduced conditions at an fO_2 close to the wustite-magnetite buffer. In addition, Mn contents of ilmenite and coexisting biotite display an apparent correlation (Fig.23) suggesting that magmatic partitioning of Mn between the two phases has been preserved.
A qualitative interpretation of the ilmenite composition data, conditional that the assumption of Mn as an inert component is valid and that ilmenites have preserved a memory of magmatic conditions, is that the ilm$_{98-100}$ compositions obtained for ilmenites from sphene-absent, high-Al biotite-bearing suites indicate these suites crystallized under reducing conditions approximated by the magnetite-wustite buffer. The ilm$_{92-95}$ compositions obtained from the sphene-bearing plutons indicate more oxidized f$_{O_2}$ conditions near the QFM and 'graphite maximum' (Ohmoto and Kerrick, 1977) curves (Fig.24). The absence of magnetite and sphene + magnetite assemblage allows an upper limit for f$_{O_2}$ conditions to be made. The West Coast Province suites were more reduced than the f$_{O_2}$ curve defined by the sphene + magnetite = hedenbergite + ilmenite + quartz reaction (Wones, in prep.). The low Fe$^{3+}$/total Fe ratios (Fig.11) of the West Coast Province suites are compatible with the ilmenite data that indicate these suites were derived from relatively reduced magmas.

Accessory minerals

Zircon and apatite are ubiquitous accessory phases observed in thin section. Primary sphene is restricted to a number of plutons in the Kulim Complex and the Bintang Batholith and its satellitic plutons. Fluorite is uncommon and only observed in rocks of leucogranitic compositions. Allanite is present in several plutons of the southern Main Range Batholith. Monazite, topaz, rutile and xenotime are commonly identified in alluvial heavy mineral concentrates by geologists of the Federal Geological Survey of Malaysia (e.g. Ng and Yong, 1969).
Fig. 24 Representative ilmenite compositional stability curves (after Spencer and Lindsley, 1981) in a $T(°C) - f_{O_2}$ plot compared to the standard hematite-magnetite, quartz-fayalite-magnetite and graphite-$CO_2$ buffer curves, the 'graphite max' curve of Ohmoto and Kerrick (1977), sphene + magnetite + quartz = hedenbergite + ilmenite reaction curve (Wones, in prep.). Stippled areas represent approximate $f_{O_2}$-$T$ ($°C$) fields of the West Coast Province granitoids indicated by ilmenite data if these are recording magmatic conditions (see text). Standard buffer curves are those for 1-2 kb.
Selected major and trace element trends of the studied West Coast Province suites displayed on conventional Harker variation diagrams are shown in Fig.25A-O. If the behaviour of trace elements in the high-SiO$_2$ leucogranites (>75% SiO$_2$) is ignored for the present moment, then patterns of trace element behaviour may be divided into four broad groups:

(a) elements that increase with increasing SiO$_2$ : U, Rb, Sn, Cs
(b) elements that decrease with increasing SiO$_2$ : CaO, Al$_2$O$_3$, FeO, MgO, TiO$_2$, P$_2$O$_5$, Zn, Zr, V, Ni, Cr, Sc, Hf, Cu, Ba, Sr, Eu
(c) elements that are decoupled with respect to SiO$_2$ : Na$_2$O, K$_2$O, Pb, Nb, Ga, Ta, REE
(d) elements that show different behaviour in different suites and suggest local fractionation control : LREE, Y, Th

In terms of the observed mineral assemblages in the West Coast Province suites, the behaviour of the group (b) elements can be qualitatively accounted for by fractionation of plagioclase, biotite and small amounts of ilmenite, zircon and apatite. Although such fractionation signatures complicate inter-batholith chemical comparisons, a number of generalized conclusions can be made with regard to actual elemental abundances. The trace element abundances of a 'typical West Coast Province granodiorite' are listed in table 4. These represent the averages of 18 samples with whole rock SiO$_2$ contents between 68% and 72%. Fig.26 compares the trace element abundances (normalized to estimates of average total crust (Taylor, 1979)) of a typical West Coast Province granodiorite with those of other granitoids from a number of batholith terrains. All the other granitoid compositions used for comparison are of roughly equivalent SiO$_2$ compositions. Fig.26 shows that, as a province, the West Coast
Fig. 25 See following page.
Fig. 25 Plots of selected trace elements against SiO$_2$ on conventional Harker variation diagrams. Concentrations along vertical axis in all plots are in ppm. Important features of these diagrams include the Rb-Ba-Sr trends and high abundances of Rb, Pb, Th, U, Sn. Note also unusually high Mn contents of some high SiO$_2$ granites.
Table 4  Trace element abundances of an average West Coast Province granodiorite - adamellite (concentrations listed represent the average of 18 samples lying in the 68-72% SiO$_2$ range).

<table>
<thead>
<tr>
<th>Element</th>
<th>Concentration (ppm ± uncertainty)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ba</td>
<td>485 ± 180 (1 cl)</td>
</tr>
<tr>
<td>Rb</td>
<td>345 ± 60</td>
</tr>
<tr>
<td>Sr</td>
<td>109 ± 39</td>
</tr>
<tr>
<td>Pb</td>
<td>50 ± 13</td>
</tr>
<tr>
<td>Th</td>
<td>47 ± 15</td>
</tr>
<tr>
<td>U</td>
<td>12.7 ± 4.5</td>
</tr>
<tr>
<td>Zr</td>
<td>203 ± 45 (excludes 1 sample with anomalously high Zr)</td>
</tr>
<tr>
<td>Nb</td>
<td>17.4 ± 3.6</td>
</tr>
<tr>
<td>Y</td>
<td>38 ± 11</td>
</tr>
<tr>
<td>La</td>
<td>39 ± 9</td>
</tr>
<tr>
<td>Ce</td>
<td>90 ± 23</td>
</tr>
<tr>
<td>Sn</td>
<td>14 ± 4.5</td>
</tr>
<tr>
<td>Sc</td>
<td>7.7 ± 1.7</td>
</tr>
<tr>
<td>V</td>
<td>33 ± 9</td>
</tr>
<tr>
<td>Cr</td>
<td>13 ± 9</td>
</tr>
<tr>
<td>Mn</td>
<td>347 ± 52</td>
</tr>
<tr>
<td>Ni</td>
<td>6 ± 3</td>
</tr>
<tr>
<td>Cu</td>
<td>5.7 ± 3.5</td>
</tr>
<tr>
<td>Zn</td>
<td>52 ± 15</td>
</tr>
<tr>
<td>Ga</td>
<td>17.5 ± 1.4</td>
</tr>
</tbody>
</table>
Fig. 26 Plot of normalized trace element abundances of a typical West Coast Province granodiorite compared with other continent margin 'orogenic granodiorites'. The typical West Coast granodiorite composition used is listed in Table 2. The normalization values used to obtain the ratios plotted are those of 'average total crust' of Taylor (1979). These values are close to those of average andesite. Data sources for the other compositions plotted are: East Coast I-type (Chapter 5); Berridale S-type (Cootralantra Granodiorite, White et al., 1977); Berridale I-type (Tara Granodiorite, White et al., 1977); Tuolumne Series granodiorite (Bateman and Chappell, 1979); average central Chilean granodiorite (Lopez-Escobar et al., 1979); Uasilau-Yau Yau (Whalen, 1980).
granitoids have high Rb, Th, U, Pb, Nb, REE, and Y compared to other orogenic granitoids. Although not shown in Fig.26, the West Coast Province also display high Sn and Cs abundances. Another notable feature of Fig.26 is the similar trace element patterns, despite differences in actual abundances, of granitoids derived from recycled old continental crust (West Coast Province, Himalayan) and granitoids located above old continental crust (Berridale, Tuolumne, Chile). The patterns of these granitoids contrast with those displayed by the New Britain examples which are associated with primitive island arc crust (Whalen, 1980; Johnson et al., 1978).

**Major element modelling**

General examination of the petrographic features and chemical data, together with consideration of likely S-type liquidus mineralogies (e.g. Clemens and Wall, 1981) indicate that plagioclase, biotite, ilmenite, zircon, apatite and possibly, quartz were early fractionating phases. The fractionation behaviour of muscovite and K-feldspar cannot as yet be ascertained but the small amounts of primary muscovite indicated based on mineral chemistry work suggests a very limited role for muscovite in the fractionation histories of the West Coast Province suites. The megacrystic habit of K-feldspar and phenocrysts of such in microgranites indicate some fractionation role for this mineral but the megacrystic habit may also reflect rapid growth rather than anything else. Only plagioclase, biotite, quartz and K-feldspar need be of concern for the major element oxide modelling in this section. The modelling will attempt to generate the general major element oxide trends from 68 to 75% SiO₂ by removal of various combinations of plagioclase, biotite, quartz and K-feldspar. The simple arithmetical relationship involved is the balancing of:
\[
\% \text{oxide in parental liquid} = \% \text{oxide in derivative liquid} + \%
\text{oxide of bulk liquidus assemblage}
\]

The behaviour of granitoid systems as predicted by crystallization relationships in the salic tetrahedron (Fig. 27) Or - Ab - An - Q may be used to describe the fractionation process modelled. Crystallization of a granitoid melt is typically initiated in the plagioclase-liquid phase volume (hypothetical point X). Biotite may accompany plagioclase. Melt composition shifts along a line XY as plagioclase crystallizes out. At point Y on the quartz saturation surface, quartz joins plagioclase (+ biotite). Quartz is interpreted as earlier than K-feldspar because quartz inclusions in the latter are a rather common feature. K-feldspar crystallization is initiated at point Z and the melt subsequently evolves along the cotectic towards the temperature minimum, with coprecipitation of plagioclase, biotite, quartz and K-feldspar.

The fractionation model discussed in this section assumes a hypothetical parental liquid with the composition of a typical 70\% SiO\textsubscript{2} granodiorite. A 70\% SiO\textsubscript{2} liquid was chosen because (1) a liquid of non-minimum melt composition is required in view of the extensive fractionation suggested by the extreme enrichment of Rb, U, Sn, etc, (2) there is a scarcity of granitoids in the West Coast Province with <70\% SiO\textsubscript{2} and (3) the chemical evolution of more siliceous parental liquids will be similar except for lower degrees of fractionation required to reach any particular derivative melt composition. Compositions of plagioclase and biotite removed at different stages of the fractionation history are based on the microprobe data discussed previously. A high-temperature alkali feldspar with 5 wt. \% Na\textsubscript{2}O was used instead of the compositions of the highly-exsolved K-feldspar megacrysts.
Fig. 27 Plagioclase-quartz - alkali feldspar crystallization relationships in the An-Ab-Qz-Or tetrahedron. Path XYZ is that typically followed by granitoid magmas. Point X is in the plagioclase + liquid phase volume - plagioclase will crystallize alone, driving the liquid composition to point Y on the quartz saturation surface. At this point, quartz and plagioclase crystallize together until point Z is reached. Plagioclase, quartz and alkali feldspar then crystallize together. Different magmas will have different crystallization paths that may be initiated and terminated at different points within this salic tetrahedron. The West Coast Province magmas, for example, seem to have initiated crystallization at the plagioclase + quartz saturation surface i.e. point Y (see text).
Although a number of fits are possible using various combinations of fractionating assemblages, the solution shown in table 5A-B are those that were reasonable in terms of petrographic evidence and also agreed with Rb - Ba - Sr modelling results discussed later. It is stressed that the results yield only a general fractionation model for evolution of granodioritic parental liquids to granitic liquids for the West Coast Province granitoid compositional range. The model is not an attempt to determine the actual liquid lines of descent or unique fractionation assemblages and does not consider complexities of restite incorporation, assimilation and variable separation of liquidus assemblages from melt. The results do, however, provide a guide to the use of reasonable proportions of fractionating phases in trace element modelling.

Several generalized conclusions can be made based on the results of the major element oxide modelling. These include:

1. Quartz and K-feldspar are required as early fractionating phases to obtain a reasonable fit to the observed trends. The achievement of quartz saturation in primitive S-type magmas indicates that quartz is a restite phase. This would mean that crystallization was not initiated in the plagioclase-liquid phase volume of the Or - Ab - An - Q tetrahedron (Fig.25) but on the quartz saturation surface, thus allowing plagioclase + liquid + quartz to coexist in the earliest magmas.

2. Granitic liquids > 75% SiO₂ can be generated after only 50 - 60% crystallization of an original 70% SiO₂ liquid. Further fractionation requires removal of solid assemblages of granite composition.
### Table 5A Fractionating phases used in major element modelling.

<table>
<thead>
<tr>
<th>Weight fraction of mineral phase removed in each 10% step*</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
</tr>
<tr>
<td>----------------</td>
</tr>
<tr>
<td>Plagioclase</td>
</tr>
<tr>
<td>Biotite</td>
</tr>
<tr>
<td>Quartz</td>
</tr>
<tr>
<td>K. feldspar</td>
</tr>
<tr>
<td>Ilmenite</td>
</tr>
</tbody>
</table>

* denotes 10% by mass of preceding liquid, not original liquid. After the seventh step, 48% by mass of original liquid remains.

### Table 5B Typical evolving liquid compositions obtained from major element oxide modelling (all values in wt. %)

<table>
<thead>
<tr>
<th></th>
<th>0</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>70.0</td>
<td>71.2</td>
<td>72.1</td>
<td>73.0</td>
<td>73.9</td>
<td>74.5</td>
<td>75.0</td>
<td>75.2</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.5</td>
<td>14.2</td>
<td>13.9</td>
<td>13.6</td>
<td>13.2</td>
<td>12.9</td>
<td>12.7</td>
<td>12.5</td>
</tr>
<tr>
<td>FeO</td>
<td>2.25</td>
<td>2.0</td>
<td>1.7</td>
<td>1.4</td>
<td>1.0</td>
<td>0.8</td>
<td>0.6</td>
<td>0.5</td>
</tr>
<tr>
<td>MgO</td>
<td>0.8</td>
<td>0.6</td>
<td>0.4</td>
<td>0.25</td>
<td>0.15</td>
<td>0.07</td>
<td>0.03</td>
<td>0.01</td>
</tr>
<tr>
<td>CaO</td>
<td>2.0</td>
<td>1.8</td>
<td>1.6</td>
<td>1.4</td>
<td>1.3</td>
<td>1.2</td>
<td>1.1</td>
<td>1.0</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.8</td>
<td>2.8</td>
<td>2.9</td>
<td>2.9</td>
<td>2.8</td>
<td>2.7</td>
<td>2.6</td>
<td>2.5</td>
</tr>
<tr>
<td>K₂O</td>
<td>5.0</td>
<td>5.1</td>
<td>5.1</td>
<td>5.1</td>
<td>5.1</td>
<td>5.1</td>
<td>5.1</td>
<td>5.2</td>
</tr>
</tbody>
</table>

### Table 5C Compositional ranges of West Coast Province granitoids at 2 wt.% SiO₂ intervals. Tabulated values may be compared with major element oxide modelling results shown in Table 5B.

<table>
<thead>
<tr>
<th></th>
<th>68-70</th>
<th>70-72</th>
<th>72-74</th>
<th>74-76</th>
<th>&gt;76</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.0-15.0</td>
<td>13.5-14.5</td>
<td>13.0-14.2</td>
<td>12.5-13.5</td>
<td>12.0-13.3</td>
</tr>
<tr>
<td>FeO</td>
<td>2.0-3.0</td>
<td>1.5-2.5</td>
<td>1.0-2.0</td>
<td>0.5-1.5</td>
<td>0.3-1.0</td>
</tr>
<tr>
<td>MgO</td>
<td>0.5-1.3</td>
<td>0.5-1.0</td>
<td>0.2-0.5</td>
<td>0.1-0.2</td>
<td>&lt;0.1-0.2</td>
</tr>
<tr>
<td>CaO</td>
<td>1.5-2.5</td>
<td>1.5-2.5</td>
<td>1.0-1.8</td>
<td>0.5-1.2</td>
<td>0.4-1.1</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.2-3.5</td>
<td>2.5-3.5</td>
<td>2.2-3.7</td>
<td>2.5-3.5</td>
<td>3.0-4.0</td>
</tr>
<tr>
<td>K₂O</td>
<td>4.5-5.5</td>
<td>4.0-6.0</td>
<td>4.8-5.5</td>
<td>4.5-5.5</td>
<td>4.5-5.3</td>
</tr>
</tbody>
</table>
3. Proportions of fractionating phases required for permissible fits vary from 30 - 50% plagioclase, 5 - 25% biotite, 20 - 30% quartz, 0 - 30% K-feldspar and minor amounts of ilmenite. The results do not strictly apply to the actinolitic hornblende-bearing plutons of the Bubu suite but the minor modal amphibole, dominance of biotite as the main ferromagnesian silicate, absence of amphibole in more felsic compositions and the similar trends for both amphibole-bearing and amphibole-absent suites indicate that amphibole did not control fractionation of the Bubu Suite plutons to any significant extent.

4. Muscovite could not have had an important role in the fractionation history of the suites studied. The high Al$_2$O$_3$ contents of muscovite would deplete liquids too rapidly in Al$_2$O$_3$ if it was a significant fractionating phase. The inference that muscovite is not an important fractionating phase is consistent with conclusions based on mineral composition data that much of the muscovite present is of subsolidus origin.

**Rb - Ba - Sr modelling**

Rb, Ba and Sr are the most useful elements for modelling fractionation processes in S-type granitoids because they are compatible trace elements in one or more of the fractionating silicate phases, plagioclase, biotite and K-feldspar. In addition, abundances of these LIL (large ion lithophile) elements are insignificant in typical accessory phases whose modal volumes are very difficult to estimate. The utility of these elements for modelling fractionation processes is further facilitated by the very large range in concentrations between the least fractionated and most fractionated
members of a suite in the West Coast Province.

The methodology adopted for Rb - Ba - Sr modelling is based on that discussed by McCarthy and Hasty (1976). The basic expression describing equilibrium partitioning behaviour is:

\[
\frac{C^L_n}{C^O} = \frac{1}{f(D-1) + 1}
\]  

(1)

where \(C^L_n\) is the concentration of a trace element in the liquid,

\(C^O\) is the original concentration of the element in the liquid,

\(f\) is the fraction of solid removed relative to the mass of the parent liquid, and

\(D\) is the bulk crystal-liquid partition coefficient.

\(D\) is calculated from the equation,

\[
D = \sum_{i=1}^{n} W_i D_i
\]  

(2)

where \(W_i\) is the weight fraction of a particular mineral phase \(i\), and

\(D_i\) is the crystal-liquid partition coefficient of the mineral phase \(i\).

For the case of incremental equilibrium crystallization where a crystal forms in equilibrium with a melt and is then prevented from further reaction, the melt composition after the \(n\)th equilibrium crystallization step is given by:

\[
C^L_n = \prod_{i=1}^{n} \left[ \frac{1}{f_i(D_i-1) + 1} \right] C^O
\]  

(3)

Di may, of course, change at different steps as ratios of various crystallizing phases change.

The corresponding concentration of a trace element of the solid assemblage at the \(n\)th increment is given by:

\[
C^S_n = \frac{D_n}{D} \cdot C^L_n
\]  

(4)
The fraction of the original liquid remaining after the \( n \)th step is obtained from the equation:

\[
f_n = (1-x)^n
\]  

(5)

where \( x \) is a constant weight fraction of the melt removed as solids at every step.

Fractionating assemblages used to calculate fractionation trends are those indicated by major element oxide modelling results. \( D \) values adopted in the computations are shown in table 6. The \( D \) values are those reported for volcanic rocks of rhyolitic and dacitic compositions and were obtained from compilations by Hanson (1978), Cox et al., (1979), Perfit et al., (1980) and Ewart (1982). It is, however, important to note that the models discussed in this section are based on idealised partitioning behaviour which can be regarded as only approximations to reality. The models do, however, serve to distinguish between contrasting igneous processes controlling chemical variations and can be used to describe the limiting effects of the various processes.

Rb - Ba - Sr interrelationships of the West Coast Province suites are shown in Fig.28 A-C. In each plot, curves representing the results of the trace element modelling calculations are shown for two hypothetical primary liquids. For each primary liquid, two curves are generated - one represents the evolution of derivative liquid compositions during the fractionation history and the other represents the compositions of the fractionated solids. Magmas that consists of liquid, liquidus crystals and entrained early fractionated solids will plot in the field bounded by the liquid and solid fractionation curves. Only values of \( f \) (fraction of primary liquid solidified) up to 0.7 are shown. This was done because (1) major element oxide results show that felsic granitic compositions can be generated after
Table 6 Partition coefficients used in trace element modelling (compiled from Hanson (1978), Cox, et al. (1979), Perfit et al., (1980) and Ewart (1982))

<table>
<thead>
<tr>
<th></th>
<th>Plagioclase</th>
<th>Biotite</th>
<th>K-feldspar</th>
<th>Quartz</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>dacite</td>
<td>rhyolite</td>
<td>dacite</td>
<td>rhyolite</td>
</tr>
<tr>
<td>Rb</td>
<td>0.05</td>
<td>0.02</td>
<td>3.3</td>
<td>2.0</td>
</tr>
<tr>
<td>Ba*</td>
<td>0.4</td>
<td>1.3</td>
<td>6.4</td>
<td>10</td>
</tr>
<tr>
<td>Sr</td>
<td>3</td>
<td>5</td>
<td>0.1</td>
<td>-</td>
</tr>
</tbody>
</table>

* The partition coefficients for Ba listed above are based on the ratios of the abundances of Ba in phenocrysts and groundmass of dacitic and rhyolitic volcanic rocks. Mittlefehldt and Miller (1982) suggest that a significantly higher partition coefficient for Ba between K-feldspar and granitic liquid of 12.9 and a lower partition coefficient for Ba between biotite and granitic liquid of 3.2 may be indicated by the experimental results of Long (1978) and their (Mittlefehldt and Miller, 1982) mineral composition data. More work is required to ascertain if these suggestions may be correct. Leeman and Phelps (1981) reported a partition coefficient of 22 ± 5 for Ba between sanidine and liquid but such elevated figures determined for anorogenic, anhydrous, high temperature rhyolites containing olivine and clinopyroxene are deemed inappropriate for use in this study.

** Actual partition coefficients used in modelling were varied between those listed for dacitic and rhyolitic compositions. Values for 'dacitic' compositions were used to model early fractionating liquids and 'rhyolitic' values were used for modelling late fractionating liquids.
Fig. 28 A-C Rb-Ba-Sr trends of the West Coast Province suites. Shown in each figure is a ruled triangular area representing model primitive liquid compositions. Crystallization and removal of plagioclase, biotite, quartz and alkali feldspar causes liquids to evolve along a liquid fractionation curve (curve marked lfc). lfc 1 and lfc 2 represent liquid fractionation curves for primitive liquid compositions denoted by 1 and 2 respectively. Arrows on lfc's indicate direction of increasing fractionation. Ticks on lfc's represent liquid compositions corresponding to 20% (by mass of remaining liquid present) incremental fractionation intervals. In terms of mass of original liquid remaining, the ticks correspond to 80%, 64%, 51%, 41% and 33% intervals. Dashed lines represent fractionation curves (sfc) of equivalent solid fractionating assemblages. Arrows and ticks on sfc's are equivalent to those of the lfc's. Fractionating magmas consisting of liquids and liquidus crystals will fall in the area bounded by the lfc and sfc curves. Calculation techniques for the incremental equilibrium fractionation mode follow those of McCarthy and Hasty (1976). Equations and the fractionating assemblages used are explained in the text. In comparison to the extended composition curves generated by fractionation, the composition curve generated by 10 to 50% partial melting (dotted curve) of a model sedimentary source is very limited - equilibrium batch partial melting alone cannot explain the large range of LIL element abundances observed in a suite and very low Sr, Ba and very high Rb abundances. Symbols denote actual measured compositions. Symbols as for Fig. 25.
lfc: liquid fractionation curve
sfc: solid fractionation curve

primitive liquid field

partial melting curve (see text)

Sr (ppm)

Ba (ppm)

Sr (ppm)

Rb (ppm)

- southern Main Range
- Penang Complex
- Bubu Suite
- Kulim Complex
The diagram shows a partial melting curve in the Rb-Ba phase space. The partial melting curve intersects with the primitive liquid field, indicating the potential for partial melting processes to occur under specific conditions. Differences in degrees of partial melting or a single source can lead to geologically important suites that yield very high Rb and very low Ba contents. This can be observed in the context of isotope geochemistry, where such suites are often associated with specific tectonic or magmatic processes.
50–60% solidification, (2) leucogranitic compositions > 75% SiO₂ show chemical signatures suggestive of processes other than simple crystal–liquid equilibrium fractionation (next section) and (3) incorporation of interstitial liquids within the framework of early crystallized solids removes significant quantities of liquids and leaves much less derivative liquid than the 30% implied by f = 0.7.

Also shown in Fig. 28 A–C is a curve generated by up to 50% partial melting of a sedimentary source with 200 ppm Sr, 160 ppm Rb, 550 ppm Ba (typical values of shales summarized in Handbook of Geochemistry – II/4). A tonalitic residue (40% plagioclase, 30% biotite, 15% quartz, 10% cordierite + garnet, 5% K-feldspar) was assumed. The advantage of showing the various melting and crystallization curves on each diagram is that it allows one to qualitatively separate the effects of melting, restite umixing (White and Chappell, 1977) and fractional crystallization processes. A number of general conclusions can be made based on the results of the trace element modelling summarized in Fig. 28 A–C:

1. Partial melting processes associated with upper amphibolite-grade metamorphism can only produce a very restricted field of magma compositions. Plutons comprising a suite cannot be produced by differences in degrees of partial melting of a single source. No geologically reasonable source can yield the very high Rb and very low Sr, Ba abundances of the rocks of leucogranitic compositions.

2. Curved trends like those displayed in the Rb vs. Sr and Rb vs. Ba plots cannot be explained by unmixing processes involving variable removal of residue and a partial melt. Unmixing processes predict straight line relationships.
3. The nature of the Rb - Ba - Sr variations can be explained by a fractional crystallization model involving variable removal of plagioclase, biotite, quartz and K-feldspar from an evolving liquid. The model suggests that the early granodiorites with low Rb and high Sr, Ba may be viewed as accumulations of early crystallizing assemblages (+ restite components) with variable incorporation of melt. Late-stage leucogranites, in contrast, represent cooled equivalents of evolved residual liquids.

4. The modelled trends indicate parental liquids whose LIL element abundances are typically 250 - 400 ppm Rb, 80 - 150 ppm Sr and 350 - 600 ppm Ba. The compositions of biotites in the most mafic granodiorites offer a means to test these estimates. Rb and Sr contents of biotites of three granodiorite samples from the Beranang Suite, Bt. Bendera Suite and the Tranum Granodiorite were available because they had been analysed for Rb - Sr dating. Their Rb contents are 1054 ppm, 1462 ppm and 733 ppm respectively whereas their Sr contents are 5 ppm, 8 ppm and 6 ppm respectively. Rb and Sr will partition between coexisting biotite and liquid according to the expression $D = \frac{\text{concentration of element in biotite}}{\text{concentration of element in liquid}}$. This expression may be used together with the biotite data for the most mafic granodiorites to obtain estimates of Rb and Sr concentrations in primitive liquids. 220 - 440 ppm Rb and 50 - 80 ppm Sr are the resultant estimates of Rb and Sr abundances of the primitive liquids using the biotite - liquid partition coefficients listed in table 6. These estimates are essentially identical to those postulated based on Rb - Ba - Sr modelling when it is considered that partition coefficients are generally difficult to determine precisely and the magnitude of variations in partition
coefficients for different rock suites are not well known.

5. The present Rb – Ba – Sr evidence suggests that suites, recognized based on structural continuity, mineralogical and chemical similarities, are produced by fractional crystallization in upper crustal chambers from a batch of magma formed during a single fusion or several closely related fusion events.

A number of qualifications, however, need to be noted with regard to the conclusions expressed above. Although it is obvious that fractional crystallization is the main mechanism controlling the chemical variations observed, it is for example impossible to distinguish closed system fractional crystallization from that which is controlled by combined fractionation and assimilation. Assimilation of sediments by an S-type magma produces signatures that cannot be rigorously identified by geochemical methods. Thus, the use of the expression describing combined assimilation-fractionation (DePaolo, 1981a) yields results that are similar to that for closed system fractionation. One line of evidence, however, may suggest that assimilation is a significant magma-modifying process. A number of plutons from the West Coast Province yield anomalously old Rb – Sr whole rock ages, from 20 to 80 Ma older than ages indicated by U-Pb zircon. Such phenomena can be plausibly attributed to assimilation (e.g. DePaolo, 1981a; James, 1982). Old apparent ages are obtained because assimilation of high $^{87}\text{Sr}/^{86}\text{Sr}$ materials can yield an alignment of points with a positive slope on an isochron diagram even before the whole rock system starts recording age information due to radioactive decay of $^{87}\text{Rb}$. 
The results of the Rb - Sr - Ba modelling argue against magma mixing during the fractionation history of a suite and also against continuous replenishment of crustal magma chambers, as both processes disrupt coherent fractionation signatures and prevent the production of highly-evolved compositions. Mixing between primitive magmas before extensive fractionation occurred cannot, however, be ruled out.

**REE**

Chondrite normalized patterns for selected samples from the West Coast Province suites are shown in Fig. 29 A-D. All patterns show LREE - enrichment and a negative Eu anomaly that increases with increasing SiO$_2$ contents. The patterns evidently reflect the REE characteristics of the sedimentary source rocks - they are essentially indistinguishable, except for Eu, from the numerous analyses of post-Archaean sediments (Taylor, 1979), with La ∼100 - 200x chondritic and Yb ∼10 - 20x chondritic. Both the LREE and HREE do not display a consistent relationship with respect to SiO$_2$, in some suites they increase, in others they decrease with increasing SiO$_2$. In a number of high-SiO$_2$ granites, a gradual depletion in La is observed. This feature is discussed in later sections. Eu shows minor variations for intra-suite samples and a general decrease with increasing fractionation. The modest magnitude of the negative anomalies can be quantitatively explained in terms of plagioclase, K-feldspar and apatite removal.

**La - Ce - Y - Th - U : monazite fractionation**

The best developed trace element signatures attributed to monazite fractionation are displayed by samples from the Beranang and Sg. Ara suites. In these suites, La, Ce, Y and Th decrease with
Table 7 REE, Cs, Hf and Ta abundances of representative samples of the West Coast Province granitoids determined by INAA (all values in ppm).

<table>
<thead>
<tr>
<th></th>
<th>Penang Complex</th>
<th>Kulim &amp; Bintang</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>La</td>
<td>47.80</td>
<td>43.90</td>
</tr>
<tr>
<td>Ce</td>
<td>102.60</td>
<td>93.90</td>
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<tr>
<td>Nd</td>
<td>42.50</td>
<td>39.50</td>
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<tr>
<td>Sm</td>
<td>8.70</td>
<td>9.00</td>
</tr>
<tr>
<td>Eu</td>
<td>0.90</td>
<td>0.90</td>
</tr>
<tr>
<td>Gd</td>
<td>7.50</td>
<td>7.00</td>
</tr>
<tr>
<td>Tb</td>
<td>1.22</td>
<td>1.37</td>
</tr>
<tr>
<td>Ho</td>
<td>1.43</td>
<td>1.61</td>
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<tr>
<td>Eu</td>
<td>3.30</td>
<td>4.70</td>
</tr>
<tr>
<td>Lu</td>
<td>0.47</td>
<td>0.68</td>
</tr>
<tr>
<td>Cs</td>
<td>17.30</td>
<td>14.30</td>
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<tr>
<td>Hf</td>
<td>6.30</td>
<td>4.70</td>
</tr>
<tr>
<td>Ta</td>
<td>2.20</td>
<td>2.20</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>Southern Main Range</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>29</td>
</tr>
<tr>
<td>La</td>
<td>31.10</td>
</tr>
<tr>
<td>Ce</td>
<td>64.40</td>
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<td>Nd</td>
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</tr>
<tr>
<td>Gd</td>
<td>4.83</td>
</tr>
<tr>
<td>Tb</td>
<td>0.90</td>
</tr>
<tr>
<td>Ho</td>
<td>1.26</td>
</tr>
<tr>
<td>Yb</td>
<td>3.04</td>
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<td>Lu</td>
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<td>Cs</td>
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<tr>
<td>Hf</td>
<td>3.60</td>
</tr>
<tr>
<td>Ta</td>
<td>3.50</td>
</tr>
</tbody>
</table>

* denotes high-SiO₂ granites
θ Genting Sempah microgranodiorite (subvolcanic unit)
Fig. 29 A-B Chondrite-normalized rare-earth element abundances of samples from the Main Range suites. Numbers in each figure represent sample numbers and SiO₂ contents (in parentheses) of the samples.
Fig. 29 C-D Chondrite-normalized rare-earth element abundances of samples from the Kulim and Penang Complexes and the Bubu Suite. Numbers in each figure represent sample numbers and SiO₂ contents (in parentheses) of the samples.
Fig. 30 La vs. Th plot for the granitoids of the West Coast Province illustrating the low La/Th ratios of the suites and the trend towards low La, low Th abundances in high SiO$_2$ granites belonging to suites that have fractionated monazite.
fractionation to very low concentrations (Fig.25). La (and Ce) shows a strong correlation with Th (Fig.30). For SiO₂ contents > 75%, samples from the Beranang and Sg. Ara Suites possess abundances of La ∼5 - 10 ppm and Th ∼10 - 15 ppm. These values contrast with those of, for example, a 76% SiO₂ leucogranite from the Ulu Kali Suite with 53 ppm Th and 51 ppm La. A plot of Y vs. La (Fig.31) shows two diverging patterns - one where Y decreases with decreasing La, which can be attributed to monazite fractionation and another where Y is decoupled from La. Fig.32 is a plot of Th vs. U showing the distinct high U - low Th field of the samples which have been interpreted to have experienced monazite fractionation. Only a (LREE + Th) - rich mineral with significant Y such as monazite can explain the consistency of the La - Th - Y trends and the decoupling of Th - U. For these reasons, it is considered here that the case for monazite fractionation is rather strong.

The La vs. Th plot of Fig.30 reveals another characteristic feature of S-type granitoids derived from sedimentary sources that have experienced considerable weathering. The West Coast Province as a whole is characterized by low La/Th ratios (< 1.5) that are independent of monazite fractionation signatures. Low La/Th ratios are also a chemical feature of recycled sediments and may plausibly be explained by the preferential adsorption of Th (and U) relative to La by clays (e.g. Bhatia and Taylor, 1981).

K-Rb ratios : K-feldspar and biotite fractionation

The West Coast Province suites display the low K/Rb ratios that are a common feature of fractionated S-types. K/Rb ratios vary from 160 for samples of granodioritic compositions to less than 50 for the most evolved leucogranites. K/Rb ratios show a slow decrease early in
Fig. 31 Plot of La vs. Y for the West Coast Province granitoids. Symbols as for Fig. 30.

Fig. 32 Plot of Th vs. U for the West Coast Province granitoids. Symbols as for Fig. 30.
Fig. 33 Variation of K/Rb ratios with SiO₂ for the West Coast Province granitoids. Most continental crustal rocks and granitoids possess K/Rb ratios of 150 - 300. There is a trend towards low K/Rb ratios with increasing fractionation but some of the high SiO₂ granites with K/Rb ratios of < 50 show features suggestive of reaction with a fluid phase. Symbols as for Fig. 30.
the fractionation history as biotite with low K/Rb ratios is removed. As the amount of biotite removed decreases and that of K-feldspar (with high K/Rb ratios) increases, the whole rock K/Rb ratios decrease to anomalously low values (Fig.33).

Mn, Sc, Zn, V, Ni, Cr, Cu

These elements are strongly partitioned into either biotite or ilmenite. Their consistent depletion with increasing fractionation and well-correlated trends with MgO and FeO can be attributed to biotite and ilmenite fractionation.

CHEMICAL VARIATIONS IN HIGH-$SiO_2$ GRANITES

A number of anomalous chemical features are displayed by analysed leucogranite samples with $SiO_2$ contents $> 75\%$ $SiO_2$. These are characterized by high Rb ($>500$ ppm), low Sr ($<50$ ppm), low Ba ($<100$ ppm) and low Zr ($<125$ ppm). Their mineralogical association is quartz + sodic plagioclase + K-feldspar + tourmaline + muscovite + garnet. For this group of samples, $SiO_2$ contents are useless as an index of degree of fractionation and samples with the highest $SiO_2$ contents may not necessarily be the most highly fractionated. Rb displays the largest range in variation but its high mobility severely restricts its reliability as an index of fractionation. Ba and Sr display little change in their very low abundances for the high-$SiO_2$ samples. Of all the other elements, only Zr displays any consistent relationship with elements like Rb, U and Sn. Zr is, in addition, a relatively immobile element. For the purposes of this study, Zr is thus possibly the best candidate for use as an approximate index of fractionation (Fig.34 A-F).
Fig. 34 A-F Plots of Pb, U, Th, Nb, Y and La vs. Zr for the West Coast Province granitoids. The large range of Zr contents (20 to 125 ppm) displayed by the high SiO₂ granites is in contrast to the small variations in SiO₂ (2-3 wt. %) contents. The trends shown in Fig. 34 suggest that differentiation processes have operated to generate large differences in trace element contents despite the limited major element oxide variations in the high SiO₂ granites. Symbols as for Fig. 30.
Rocks of the high-SiO₂ granites often exhibit distinctive features that are not present in low-SiO₂ granites.

Fig. 34 See previous page
Rocks of the high-SiO₂ leucogranite association display two distinctive features that set them apart from more typical granitoids with which they are associated. The first is the rapid increases in abundances of elements like Rb, U, Sn and Cs (Fig. 25) with little change in major element chemistry and the second is the break from the main chemical trends defined by the less differentiated granodiorite - granite suites for elements like Nb, Y, Ga, Mn and others. (Fig. 34, 35). Elements that are enriched in the high-SiO₂ granites are Rb, U, Sn, Cs, Na, P, Nb, Ga, Mn and almost certainly Li, Be, B, F. Elements that are depleted include Ni, Cr, V, Cu, Zn, Sc, Pb, Y, LREE, Eu and Th. Fig. 35 A-B shows the anomalous fields of a number of the high-SiO₂ granites on plots of Ga vs. Al₂O₃ and Zn vs. Mn. Such plots show that the compositions of these high-SiO₂ leucogranites are completely at variance with those expected if fractional crystallization was the only process controlling chemical differentiation.

REE patterns of the high-SiO₂ leucogranites are shown in Fig. 36. Two groups can be recognized from Fig. 36 - a first group with LREE enrichment, modest negative Eu anomalies and slightly higher than average HREE contents, and a second group with extremely depleted and erratic LREE patterns, extreme negative Eu anomalies and variable HREE contents. Although the behaviour of the LREE elements may be attributed to monazite fractionation, the magnitude of the negative Eu anomalies of the second group cannot be quantitatively explained by feldspar fractionation using presently known partition coefficients.

The chemical features described above suggest that mechanisms other than simple crystal-liquid equilibrium fractional crystallization need to be invoked to describe the petrogenesis of the
Fig. 35A. Plot of Ga vs. $\%\text{Al}_2\text{O}_3$ illustrating departures from the main trend shown by some high SiO$_2$ granites. Symbols as for Fig. 30.

Fig. 35B. Plot of Zn vs. Mn illustrating the anomalous high Mn contents of some high SiO$_2$ granites that deviate from the main trend defined by less felsic granitoid compositions. Symbols as for Fig. 30.
Fig. 36 Chondrite-normalized rare-earth element abundances of several representative high SiO₂ granites. Samples represented by solid symbols show anomalous Ga, Zn abundances (Fig. 35, 36). These also show extreme negative Eu anomalies and erratic rare-earth element patterns. Such features are best explained by interaction with a fluid phase. Sample numbers and SiO₂ contents (in parentheses) are shown at bottom right-hand corner of figure.
high-SiO₂ leucogranites.

Evolution of a fluid phase and infiltration metasomatism

Jahns and Burnham (1969) and Burnham (1978, 1980) have discussed the mechanisms involved in second boiling and evolution of a separate fluid phase from magmas that are water-saturated. Under different physical conditions, different fluid phases will coexist with the parent silicate magma. At high temperatures, the fluid will be gaseous. Elements in the melt + crystal + volatile phase (gas?) system will partition "such that the chemical potential of each chemical species is the same in all phases at equilibrium" (Burnham, 1980). Volatiles like H₂O, CO₂, F⁻, Cl⁻, BO₃⁻ etc are preferentially partitioned into the volatile phase together with elements like Li, Be, Sn, Ta, Nb that are thought to form chloride and fluoride complexes. Pegmatite and aplite masses result from the cooling of the volatile phase. The degree of complexity of the pegmatite-aplite-granite association will be related to the extent of the complex interaction and replacement processes established between the melt phase and volatile phases.

The ubiquity of pegmatite – aplite bodies associated with the West Coast Province suites shows that attainment of water-saturated conditions, and generation and expulsion of a separate fluid phase was a widespread phenomenon. Three phases can thus be recognized during the late-stage evolution of the granitic magmas – a water-saturated, highly fractionated silicate melt (+ crystals), a residual silicate melt (+ crystals) from which a volatile phase has evolved and lastly, the volatile phase itself. Interactions between the various phases are likely and offer a framework to discuss the chemical signatures of the leucogranites. A major problem facing such an attempt is the
difficulty in distinguishing between signatures inherited from a long history of fractional crystallization and plausible liquid state differentiation processes like volatile complexing from signatures produced by reaction with a fluid phase.

The low abundances of elements like Ba, Sr (plagioclase + K-feldspar fractionation) and V, Ni, Cr, Sc (biotite and ilmenite fractionation) can be unambiguously attributed to fractional crystallization. Features like modest negative Eu anomalies, high Rb, U, Sn, Cs and low La, Ce, Th must at least, in part, be the result of fractional crystallization. A number of leucogranites display the characteristics listed above and are suspected to represent magmas that did not reach the critical stage of fluid separation. Other leucogranites however display extreme negative Eu anomalies, very high Rb (> 600 - 700 ppm), Sn and anomalous abundances of Ga, Mn, P, Na₂O, Nb. These samples are suspected to have, in some way, reacted with late-stage fluids. Enrichment of Ga, Mn, P, Na₂O, Nb in addition to high Rb, U, Sn, Cs are best explained by a process of local late-stage infiltration metasomatism superimposed over fractionation signatures. Such a mechanism would be compatible with the deduced late-stage textural and chemical features of associated tourmaline, muscovite, spessartine and albite. The most obvious metasomatic texture is that of the interstitial skeletal network of late tourmaline in many leucogranites.

Volatile complexing

The process of infiltration metasomatism by late-stage volatile-enriched fluids of magmatic origin are believed to be responsible for the anomalous enrichment of elements like Rb, Sn, Ga, Mn, P, Na₂O, Nb. Preferential enrichment of such elements may be
attributed to the enhanced transport of these elements as molecular complexes in late-stage fluids (e.g. Burnham, 1980; Collins et al., 1982 and references cited therein). The special role of chloride and fluoride complexes have been stressed by many workers and contrasts with the generally lesser emphasis given to the role of CO$_2$- and BO$_3$-complexes. Chloride complexes have been favoured because chloride minerals are not stable in silicate melts and, thus, chlorine is preferentially partitioned into the fluid phase. In addition, chloride complexes are highly stable in aqueous fluids at magmatic temperatures (Burnham, 1979). An important role is indicated for fluorine by the work of Ishihara et al., (1980). These workers obtained F and Cl analyses for granitoid samples in Thailand that are northern extensions of the West Coast Province batholiths. These Thai granitoids possess unusually high F contents and exceptionally high F/Cl ratios. F contents average ≈ 1700 ppm (400 - 6000 ppm) and can be compared with values of ≈ 800 ppm for typical granitoids (Handbook of Geochemistry II/1, II/2). Similarly, their F/Cl ratios of 10 - 100 are markedly higher than values of 3 - 8 for typical granitoids.

The formation of stable chloride complexes can explain the high Mn and slightly elevated Zn contents of leucogranites that are interpreted to have reacted with a fluid phase (Fig. 35 B). Holland (1972) documented extensive partitioning of Mn and Zn towards an aqueous chloride phase, perhaps as complexes of the MnCl$_2$-type. Sn$^{2+}$ may similarly form chloride complexes (SnCl$_2$$_4$-, SnCl$_2$$_2$) and be preferentially enriched in the fluid phase.

Enrichment of Ga, Nb and Sn in a fluid phase may be attributable to fluorine complexing. Eadington and Giblin (1979) and Collins et al., (1982) noted that SnF$_4$+, GaF$_6$-, (NbF$_5$)$_2$, Na$_2$NbF$_7$,
SnF$_4$ and other complexes are possibly stable in a volatile-enriched magmatic-hydrothermal fluid.

Volatile complexing may occur in both the silicate melt as well as the fluid phase. It is clear however that greatly enhanced rates of complexing will be displayed in a volatile-rich fluid phase.

**Behaviour of REE in leucogranites**

Although enrichment patterns are quite adequately explained by reaction with a volatile fluid phase, the depletion patterns of the LREE and in particular, Eu are more difficult to explain. The La (and Ce) – Th correlation and the decoupling of U and Th were explained in terms of monazite fractionation in a previous section. Although monazite fractionation can explain the low La, Ce abundances, low LREE abundances may, however, also be a subsidiary effect of reaction with a fluid phase. The best documented evidence for this comes from the work of Alderton et al. (1980), who showed that late-stage tourmalinization is accompanied by marked reduction in LREE contents. More significantly, breakdown of plagioclase and K-feldspar to muscovite accompanying fluid interaction results in significant decreases in Eu due to the absence of suitable sites in muscovite to accommodate the released Eu. The released Eu and LREE may be transported away as fluoride, chloride or carbonate complexes. Flynn and Burnham (1978) showed that Cl- and F-rich vapour phases can transport REE although the amounts involved are uncertain. In the case of the West Coast Province leucogranites, fluid phase interaction is believed to have merely amplified LREE and Eu depletions inherited from a long history of fractional crystallization.
Enrichment of Sn

Sn, together with Rb, U, Cs, displays a marked increase during magmatic differentiation from granodiorite to leucogranite compositions. Granodiorites and mafic granodiorites have abundances of 10 - 15 ppm Sn whereas abundances of 30 - 60 ppm are displayed by leucogranites (Fig. 37). Samples with the highest Sn abundances are invariably those that show anomalous enrichment of Ga, Nb, Mn and these samples have been suggested previously as having reacted with a magmatic fluid phase. Although magmas with high Sn contents like those of the West Coast Province and extended fractionation histories are conducive for Sn enrichment in residual melts, the most significant Sn enrichment process is the generation of the volatile and rare metal-rich fluid phase. Extended fractionation increased Sn contents from initial abundances of 10 - 15 ppm to 20 - 25 ppm i.e. by about a factor of two. Reaction with a fluid phase and attendant infiltration metasomatism further increased Sn abundances up to at least 55 ppm.

Sn enrichment during magmatic differentiation has been suitably explained by numerous past studies (e.g. Ringwood, 1955; Ishihara, 1979). Sn may occur in silicate melts as SnO$_4^{-}$, Sn$^{2+}$ and Sn$^{4+}$. Sn$^{4+}$ has an ionic radius similar to that of Ti$^{4+}$ and Fe$^{3+}$ and may substitute for these elements in sphene, biotite and ilmenite. In reduced magmas, however, Sn$^{2+}$ probably exists as the dominant species. Both Sn$^{2+}$ and SnO$_4^{-}$ do not enter into rock forming minerals and may be considerably enriched in highly differentiated low fO$_2$ granitic liquids. Subsequent volatile complexing during the generation of a separate fluid phase allows Sn to be preferentially concentrated in the fluid phase. To attain water-saturated conditions
Fig. 37 Variation of Sn with Rb in the West Coast Province granitoids. Samples with Rb > 500 ppm and Sn > 30 ppm are high SiO₂ granites.
necessary for generation of a separate fluid phase, either high initial water contents or extensive removal of anhydrous phases that allow water contents to build up or both are required.

The above mechanisms of Sn enrichment in granitoid magmas indicate that late-stage magmatic processes seem to be more important than original Sn abundances in determining which granitoid suites are possibly associated with tin mineralization and which are not.

COEVAL VOLCANISM

Felsic volcanic and subvolcanic rocks dated by U-Pb zircon, Rb - Sr whole rock and K - Ar biotite methods (discussed in chapter three) as late Triassic in age occur as units unconformably associated with Palaeozoic metasediments in a sedimentary screen that separates the Beranang and Ulu Kali Suites of the Main Range Batholith. The volcanic rocks are of rhyodacite - rhyolite composition with SiO₂ contents ranging from 70 to 73% (MAL 73 - 77, appendix 6). A microgranodiorite unit of subvolcanic character is associated with the volcanic rocks and this unit has SiO contents of 66 - 68% SiO₂ (MAL 78 - 80, appendix 6).

The volcanic rocks are porphyritic and contain phenocrysts of biotite, quartz, andesine and microperthitic K-feldspar. The groundmass is devitrified but recrystallized glass shards can be discerned. The groundmass of the microgranodiorite unit is coarser and in addition to biotite - quartz - K-feldspar, phenocrysts of hypersthene and labradorite - andesine are present. Phenocrysts of labradorite and hypersthene generally occur as large clots and occasionally as granoblastic aggregates together with cordierite, quartz and biotite. In these aggregates, cordierite has Mg/(Mg + Fe)
values of 0.67 to 0.69, coexisting biotite has values of 0.56 to 0.59 and coexisting orthopyroxene has values of 0.47 to 0.50. This sequence of \( \frac{Mg}{(Mg + Fe)} \) values i.e. \( m_{opx} < m_{bio} < m_{cord} \) is typical of that for pelitic granulites and this argument, together with the granoblastic textures of the orthopyroxene - cordierite - biotite - quartz - labradorite aggregates suggest that such aggregates may be regarded as fragments of high-grade metamorphic rocks. Wyborn et al., (1981) reported similar occurrences of phenocryst and xenolith assemblages in an S-type volcanic suite in southeastern Australia. They interpreted such assemblages to represent granulite - grade fragments from the source regions of the volcanics.

Both the volcanic and subvolcanic units display chemical features that are similar to the granitoid suites. They are peraluminous and have high Rb/Sr ratios, high Pb, Th, U, Sn and high \( K_2O/Na_2O \).

Initial \( \frac{^{87}Sr}{^{86}Sr} \) ratios of \( \approx 0.726 \) (chapter three) for these units are high and in the same range as the granitoid suites of the West Coast Province.

**ISOTOPIC CHARACTERISTICS OF THE WEST COAST PROVINCE GRANITOIDS**

**Sr initial ratios**

Sr initial ratios of the suites studied are high (> 0.710) and typical of granitoids derived from metapelitic sources. Precise initial ratios are, however, difficult to determine. This arises from a number of factors. Firstly, a large number of Rb - Sr whole rock determinations on samples from individual suites indicate that these suites often are characterized by variable initial ratios. Secondly, some suites have Rb - Sr whole rock ages that are anomalously old
compared with U-Pb zircon and K-Ar mica ages. Arrays defining such old ages may be regarded as inherited isochrons, following Roddick and Compston (1977). One way to generate inherited isochrons is by assimilation of highly radiogenic country rocks. Such a process can yield a crude alignment of points (assuming a simple two end-member mixing model between magma and country rocks and disregarding more complex effects that may be generated when simultaneous assimilation and fractionation occurs) between those for the granitoid magma and the assimilated components on an isochron diagram. In such a case, age-corrected initial ratios deduced from present-day measurements will not record a unique value but will lie in a range between the two end-member components i.e. between the initial ratio of the original granitoid magma and the isotopic ratio of the assimilated components at the time of intrusion. The initial ratios of the suites listed in table 8 are mainly derived from intersections of isochrons defined by 3 or more points. This was felt to be the most useful manner of listing initial ratios even when the isochrons gave anomalously old ages. In such cases, the isochron intersections probably yield the best available estimates of the intrinsic initial ratio of the granitoid magma before increases occurred due to assimilation. The disadvantage is that such initial ratios are model dependent. The values do, however, yield minimum estimates of initial ratios even if the assimilation model suggested is erroneous.

Sr initial ratios listed in table 8 range from 0.711 to 0.732. One pluton of uncertain association for which a new Rb-Sr determination is available yields an initial ratio of 0.748 if corrected using the U-Pb zircon age. Consideration of the other data points of Bignell and Snelling (1977) for samples from the same locality however suggests open system behaviour for the Rb-Sr whole
Table 8 Sr initial ratios of the studied West Coast Province suites.

<table>
<thead>
<tr>
<th>Suite</th>
<th>Initial ratio</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Southern Main Range Batholith</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Rembau Suite</td>
<td>0.710 ± 0.002</td>
<td></td>
</tr>
<tr>
<td>2. Beranang Suite</td>
<td>variable, 0.720 - 0.732</td>
<td></td>
</tr>
<tr>
<td>3. Ulu Kali Suite</td>
<td>0.711 ± 0.001</td>
<td></td>
</tr>
<tr>
<td>4. Unnamed pluton of uncertain</td>
<td>0.748 (?)*</td>
<td></td>
</tr>
<tr>
<td>association east of Ulu Kali</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5. Tranum Granodiorite</td>
<td>0.722 ± 0.002</td>
<td></td>
</tr>
<tr>
<td>6. Genting Sempah volcanic and</td>
<td>0.726 ± 0.002</td>
<td></td>
</tr>
<tr>
<td>subvolcanic units</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Southern lobe of Bintang</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Batholith</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Bubu Suite</td>
<td>0.723 ± 0.003</td>
<td></td>
</tr>
<tr>
<td>2. Unnamed pluton along southern margin of Bubu Suite</td>
<td>&lt;0.720 (?)**</td>
<td></td>
</tr>
<tr>
<td>Kulim Complex</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Penanti Suite</td>
<td>0.712 ± 0.003</td>
<td></td>
</tr>
<tr>
<td>2. Bongsu Suite</td>
<td>0.721 ± 0.002</td>
<td></td>
</tr>
<tr>
<td>Penang Complex</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. St. Bendera Suite</td>
<td>variable, 0.710 - 0.724</td>
<td></td>
</tr>
<tr>
<td>2. Sg. Ara Suite</td>
<td>0.711 ± 0.003</td>
<td></td>
</tr>
</tbody>
</table>

* initial ratio deduced from 1 sample corrected using U-Pb zircon age
** only one sample analysed. Initial ratio is that assuming an age of 200-210 Ma.
rock system. The initial ratio suggested for this particular pluton can be regarded only as tentative until more data become available.

Nd initial ratios

Nd initial ratios were calculated from present-day measured ratios by correcting for radioactive decay since the time of crystallization using ages determined by the U-Pb zircon method. If these were not available, Rb-Sr whole rock isochron and K-Ar mica ages were used where both ages showed general agreement. Initial Nd ratios are, however, relatively insensitive to errors in ages of 10 - 20 Ma because of the long half-life of $^{147}\text{Sm}$. Initial ratios are expressed here in the convenient $\varepsilon$-notation following DePaolo and Wasserburg (1979). Initial $\varepsilon_{\text{Nd}}$ is defined as:

$$
\varepsilon_{\text{Nd}} = \left[ \frac{\left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_\text{sample}^{\text{T}}}{\left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_\text{CHUR}^{\text{T}}} - 1 \right] \times 10^4
$$

where $T$ = age of crystallization

$$
\left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_\text{sample}^{\text{T}} = \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_\text{meas}^{\text{T}} - \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_\text{meas}^{\text{T}} (\lambda_{\text{Sm}} T - 1)
$$

$$
\left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_\text{CHUR}^{\text{T}} = \left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_\text{CHUR}^{\text{0}} - \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_\text{CHUR}^{\text{0}} (\lambda_{\text{Sm}} T - 1)
$$

$$
\left( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_\text{CHUR}^{\text{0}} = 0.511836 \quad ; \quad \left( \frac{^{147}\text{Sm}}{^{144}\text{Nd}} \right)_\text{CHUR}^{\text{0}} = 0.1967
$$

$\lambda_{\text{Sm}} = 6.54 \times 10^{-12} \text{yr}^{-1}$

and CHUR values denote the commonly accepted values for a reference 'chondritic uniform reservoir'.

Table 9 lists the $\varepsilon_{\text{Nd}}$ values of a number of representative samples of the West Coast Province suites together with other pertinent data required for calculation of initial $\varepsilon_{\text{Nd}}$ values. The samples show values ranging from -6 to -10. Such negative values are those expected for magmas derived from partial melting of old crust and are consistent with the high Sr initial ratios.
Table 9 Nd isotopic data for representative samples of the West Coast Province granitoid suites and coeval volcanics.

<table>
<thead>
<tr>
<th>Sample/suite</th>
<th>Age correction (Ma)</th>
<th>ppm</th>
<th>147Sm/144Nd</th>
<th>143Nd/144Nd meas</th>
<th>143Nd/144Nd T</th>
<th>εNd</th>
</tr>
</thead>
<tbody>
<tr>
<td>13/Sg. Ara</td>
<td>210</td>
<td>7.82</td>
<td>36.93</td>
<td>0.1281</td>
<td>0.51124</td>
<td>-9.8</td>
</tr>
<tr>
<td>14/Bl. Bendera</td>
<td>210</td>
<td>8.45</td>
<td>39.89</td>
<td>0.1280</td>
<td>0.51135</td>
<td>-7.6</td>
</tr>
<tr>
<td>20/Penanti</td>
<td>200</td>
<td>6.69</td>
<td>34.64</td>
<td>0.1168</td>
<td>0.51131</td>
<td>-8.1</td>
</tr>
<tr>
<td>27/Bubu</td>
<td>215</td>
<td>6.23</td>
<td>37.72</td>
<td>0.1358</td>
<td>0.51131</td>
<td>-9.6</td>
</tr>
<tr>
<td>29/Berangang</td>
<td>200</td>
<td>15.91</td>
<td>68.91</td>
<td>0.1397</td>
<td>0.51143</td>
<td>-8.9</td>
</tr>
<tr>
<td>45/Tranum</td>
<td>210</td>
<td>6.67</td>
<td>20.60</td>
<td>0.1301</td>
<td>0.51126</td>
<td>-9.8</td>
</tr>
<tr>
<td>73/Volcanic</td>
<td>215</td>
<td>6.14</td>
<td>28.54</td>
<td>0.1130</td>
<td>0.51133</td>
<td>-7.9</td>
</tr>
<tr>
<td>78/Volcanic</td>
<td>215</td>
<td>8.89</td>
<td>46.66</td>
<td>0.1153</td>
<td>0.51133</td>
<td>-7.6</td>
</tr>
</tbody>
</table>

Table 10 Sr isotopic data for representative samples of the West Coast Province granitoid suites and coeval volcanics.

<table>
<thead>
<tr>
<th>Sample/suite</th>
<th>Age correction (Ma)</th>
<th>ppm</th>
<th>87Rb/86Sr</th>
<th>(87Sr/86Sr) meas</th>
<th>(87Sr/86Sr) T</th>
<th>εSr</th>
</tr>
</thead>
<tbody>
<tr>
<td>13/Sg. Ara</td>
<td>210</td>
<td>62</td>
<td>409</td>
<td>19.3</td>
<td>0.7924</td>
<td>+4.30</td>
</tr>
<tr>
<td>14/Bl. Bendera</td>
<td>210</td>
<td>74</td>
<td>306</td>
<td>12.0</td>
<td>0.7600</td>
<td>+2.78</td>
</tr>
<tr>
<td>20/Penanti</td>
<td>210</td>
<td>79</td>
<td>44.7</td>
<td>16.5</td>
<td>0.7652</td>
<td>+2.78</td>
</tr>
<tr>
<td>27/Bubu</td>
<td>200</td>
<td>244</td>
<td>306</td>
<td>3.7</td>
<td>0.7328</td>
<td>-4.50</td>
</tr>
<tr>
<td>29/Berangang</td>
<td>215</td>
<td>40</td>
<td>432</td>
<td>31.2</td>
<td>0.8272</td>
<td>+3.88</td>
</tr>
<tr>
<td>38/Unnamed</td>
<td>200</td>
<td>30</td>
<td>643</td>
<td>64.2</td>
<td>0.9308</td>
<td>+6.24</td>
</tr>
<tr>
<td>45/Tranum</td>
<td>210</td>
<td>51</td>
<td>319</td>
<td>18.1</td>
<td>0.7824</td>
<td>+3.40</td>
</tr>
<tr>
<td>73/Volcanic</td>
<td>215</td>
<td>67</td>
<td>353</td>
<td>15.3</td>
<td>0.7706</td>
<td>+2.76</td>
</tr>
<tr>
<td>78/Volcanic</td>
<td>215</td>
<td>146</td>
<td>254</td>
<td>5.0</td>
<td>0.7415</td>
<td>+3.07</td>
</tr>
</tbody>
</table>
\( \varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}} \) relationships

As for Nd initial ratios, Sr initial ratios can also be expressed in the \( \varepsilon \)-notation using the expression:

\[
\varepsilon_{\text{Sr}} = \left[ \frac{(\frac{^{87}\text{Sr}}{^{86}\text{Sr}})^{T}_{\text{sample}}}{(\frac{^{87}\text{Sr}}{^{86}\text{Sr}})^{T}_{\text{CHUR}}} - 1 \right] \cdot 10^4
\]

where \( T \) = crystallization age

\[
(\frac{^{87}\text{Sr}}{^{86}\text{Sr}})^{T}_{\text{sample}} = (\frac{^{87}\text{Sr}}{^{86}\text{Sr}})^{T}_{\text{meas}} - (\frac{^{87}\text{Rb}}{^{86}\text{Sr}})^{0}_{\text{meas}} (e^{\lambda T} - 1)
\]

\[
(\frac{^{87}\text{Sr}}{^{86}\text{Sr}})^{0}_{\text{CHUR}} = 0.7048 \quad ; \quad (\frac{^{87}\text{Rb}}{^{86}\text{Sr}})^{0}_{\text{CHUR}} = 0.08
\]

\( \lambda_{\text{Rb}} = 1.42 \times 10^{-11} \text{ yr}^{-1} \)

Table 10 lists the pertinent data used to calculate \( \varepsilon_{\text{Sr}} \) values for samples which have also been analysed for Nd.

\( \varepsilon_{\text{Sr}} \) values calculated as above are plotted against \( \varepsilon_{\text{Nd}} \) values in Fig. 38. In this \( \varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}} \) plot, the West Coast Province samples define a subhorizontal array. Subhorizontal patterns appear to be typical of S-type granitoids derived from old crust (e.g. McCulloch and Chappell, 1982). Such patterns can be explained in terms of the contrasting behaviour of REE and Rb-Sr systems during sedimentary reworking processes. In the sedimentary environment, Sm and Nd are not significantly fractionated whereas sediments are enriched in Rb relative to Sr. This leads to rapid evolution of \( \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \) but no significant enhancement of \( \frac{^{143}\text{Nd}}{^{144}\text{Nd}} \) evolution. Consequently, \( \varepsilon_{\text{Nd}} \) of S-type granitoids will reflect the Nd isotopic characteristics of pre-sedimentary stage protoliths whereas \( \varepsilon_{\text{Sr}} \) will reflect the extent of the sedimentary protolith history.

The variable and high \( \varepsilon_{\text{Sr}} \) values of the West Coast Province may, thus, be interpreted as the result of derivation from high Rb/Sr sedimentary sources that have been extensively weathered and recycled.
Fig. 38 $\varepsilon_{\text{Nd}}$ vs. $\varepsilon_{\text{Sr}}$ diagram for the West Coast Province granitoids. Initial Nd and Sr ratios expressed in notation are obtained by calculations explained in text and listed in tables 8 and 9. The 'mantle array' defined by MORB's, ocean island basalts and some continental flood basalts is after DePaolo and Wasserburg (1979). The calculation of $T_{\text{DM}}^{\text{Nd}}$ ages on the right hand vertical axis is explained in chapter 6 and is based on the evolution of 'typical continental crust' with a $^{147}\text{Sm}/^{144}\text{Nd}$ value of 0.12.
The narrow range of $\varepsilon_{Nd}$ values suggests that these have not been significantly changed by upper crustal processes and may be used to estimate the duration of crustal prehistory. An $\varepsilon_{Nd}$ range of $-6$ to $-10$ for 200 Ma granitoid rocks is the range that would be expected for juvenile (i.e. first stage mantle derived) crust that has evolved or 'aged' for about 1500 Ma (chapter 6). The Nd isotopic data thus suggest pre-sedimentary crustal protoliths of mid-Proterozoic age for the West Coast Province granitoids.

**U-Pb zircon inheritance**

The phenomenon of zircon inheritance which describes survival of much older components in young zircon systems has been widely reported in granitoids (e.g. Pidgeon and Aftalion, 1978; Williams, 1977). Zircon inheritance is a particularly common feature shown by S-type granitoids and this is probably due to the common occurrence of detrital zircon grains in sediments. In addition, the generally low temperatures ($\sim 700^\circ$C) required to generate S-type magmas are believed to be insufficient to cause total resetting or complete dissolution-breakdown of old zircons in the source regions of the granitoids (chapter three).

The widespread zircon inheritance features documented in chapter three are thus compatible with the S-type nature of the West Coast Province suites. More importantly, zircon inheritance allow estimates of the ages of the old incorporated zircons to be made. Such inheritance ages reflect the ages of the ultimate igneous protoliths from which the old zircons were derived. U-Pb zircon results discussed in chapter three indicates inherited zircon components with $\sim 1500 - 1700$ Ma ages.
SUMMARY

Overall S-type characteristics

Granitoid suites in the West Coast Province of Peninsular Malaysia can be distinguished as S-type granitoids derived from melting of metasedimentary crust. Considerable variations exist between suites. Two broad groupings of the West Coast Province granitoids can be discerned - one group is characterized by the mafic mineral assemblage: high-Al biotite + muscovite + spessartine garnet and the second, less voluminous group has the mafic assemblage: low-Al biotite + sphene + actinolitic hornblende. Such assemblages contrast with the cordierite + Fe-Mg garnet + sillimanite - bearing granitoid suites of southeastern Australia (Hine et al., 1978; Phillips et al., 1981).

Geochemically, the West Coast Province suites are only moderately peraluminous with mol. \( \frac{Al_2O_3}{(Na_2O + K_2O + CaO)} \) ranging from 1.03 to 1.23. This feature together with high \( Na_2O \) contents (2.5 - 3.5 wt. %) and high \( K_2O \) contents (4.5 - 5.5 wt. %) are notable differences compared with the Australian Berridale-Kosciusko S-types. Chemical features characteristic of S-type granitoid suites include high \( K_2O/Na_2O \), low \( CaO \), low \( Fe^{3+}/total \ Fe \), restricted \( SiO_2 \) range of individual suites, abundance of high \( SiO_2 \) compositions, high \( Rb, Th, U, Pb, Y, Cs, Sn \) and high \( Sr \) initial ratios.

Composite batholiths and suite recognition

The granitoid batholiths and complexes of the West Coast Province can be subdivided into suites in a manner analogous to Cordilleran I-type batholiths (e.g. Cobbing and Pitcher, 1972). Suites contain a variety of rock types ranging from granodiorite through to granite,
aplite and pegmatite compositions. Samples from a single suite can be shown to be related to each other primarily by a crystal fractionation mechanism.

**Differentiation processes**

Major and trace element geochemical trends can be modelled by closed system crystal fractionation, but results are not significantly different if a combined fractionation-assimilation mode is assumed. Conventional crystal fractionation processes + assimilation cannot explain the extreme enrichment of elements like Rb and Sn, anomalous abundances of Mn, Ga, Nb and extreme depletion of Eu in a number of highly differentiated leucogranites. Such signatures are best explained by metasomatic reactions with a fluid phase generated by second boiling of water-saturated magmas. Volatile complexing may have been important in enriching certain elements during the magmatic stage but was certainly instrumental in enriching Mn, Ga, Nb, Sn in the fluid phase. Although originally high Sn abundances of the West Coast Province magmas and their extended fractionation histories are important in increasing Sn contents in residual liquids, separation of a Sn-rich, volatile-rich fluid phase is regarded as the critical factor that eventually decides whether economic concentrations of Sn are attained.

**Isotopic characteristics**

High Sr initial ratios (> 0.710) of the West Coast Province suites are consistent with their metasedimentary origin. Negative $\varepsilon_{Nd}$ initials of -6 to -10 and inherited zircon ages of $\sim 1500 - 1700$ Ma suggest that the metasedimentary sources of the granitoids were derived from ultimate igneous protoliths of mid-Proterozoic age. On
an $\varepsilon_{Nd} - \varepsilon_{Sr}$ plot, the West Coast Province samples define a
subhorizontal pattern similar to other S-type batholiths derived from
recycled old Precambrian continental crust.

**Magma generation**

The mechanisms involved in and and heat sources for generation of
batholithic volumes of S-type granitoid magma are often difficult to
explain. Mantle-derived basaltic magmas that intrude and become
intimately lodged into lower crustal rocks causing widespread melting
are often called upon as the 'heat trigger' in calcalkaline provinces
(e.g. Hildreth, 1981). This mechanism cannot be invoked for the
exclusively S-type West Coast Province where there is no evidence for
coeval mafic plutonism or volcanism. Thompson (1982) suggested that
large volumes of granitoid magma derived from anatexis of pelitic
sequences may be generated when there is an overlap of P-T conditions
for both mica dehydration and melting reactions. Three simplified
dehydration-melting reactions as discussed by Clemens and Wall (1981)
may be considered.

1. At relatively low temperatures of 650 - 700°C and 4 - 7 kb
pressures (kyanite-sillimanite facies), maximum overlap of
muscovite dehydration and initial melting of pelitic compositions
takes place via reactions of the type:

\[
\text{Muscovite} + \text{quartz} + \text{plagioclase} = \text{K-feldspar} + \text{aluminosilicate} + \text{melt}
\]

Granitic melts produced are strongly peraluminous and water-rich
(even before extensive removal of anhydrous phases by
fractionation). Such hydrous, low temperature melts do not move
far from their sites of generation before solidifying. Muscovite
is early and is probably the dominant mica. An example of this
'low-temperature' granitoid group may be the Himalayan Tertiary leucogranites (Dietrich and Gansser, 1982).

2. At slightly elevated temperatures (~700 - 750°C), Fe-biotite dehydration overlaps with melting at 5-8 kb to produce moderately peraluminous melts via reactions of the type:

\[
\text{biotite} + \text{aluminosilicate} + \text{quartz} + \text{feldspars} = \text{K-feldspar} + \\
\text{garnet} + \text{cordierite} + \text{melt}
\]

Melts formed are water-undersaturated but may contain relatively high water contents of 6-8 wt.% (Clemens and Wall, 1981). Initial liquids may not be of minimum melt compositions and they have the potential to undergo extensive fractionation to achieve water saturation.

3. At high temperatures in excess of ~750 - 800°C, small volumes of peraluminous magma may be generated by reactions like:

\[
\text{biotite} + \text{quartz} + \text{feldspars} = \text{orthopyroxene} + \text{melt} + \\
\text{feldspars} + \text{garnet} + \text{cordierite}
\]

Melts derived by such reactions are very water-undersaturated. They may ascend to shallow crustal depths or erupt as S-type volcanics. Clemens and Wall (1981) and Wyborn et al., (1981) described volcanics that may have originated in this fashion. Minor occurrences of rhyolitic - dacitic volcanic and subvolcanic sequences containing occasional orthopyroxene + cordierite + plagioclase + quartz clots in the West Coast Province are also best explained as resulting from reactions like (3) above.

Complexities observed for magma generation and modifying processes often make it difficult to describe magma generation in terms of simple single stage equilibria as discussed above.
Nevertheless, the moderately peraluminous nature of the West Coast Province granitoids, the general paucity of early primary muscovite, relatively high water contents and deduced fractionation histories indicate that their mode of generation is best described by reactions like (2) above. One way to explain the voluminous and exclusively S-type nature of the West Coast Province suites without resorting to external 'heat triggers' of mantle origin is to underthrust upper crustal metapelitic sequences to the great depths that are required for dehydration-melting under kyanite-sillimanite facies P-T conditions. A continental collision origin for the West Coast Province granitoids as suggested by Mitchell (1977) is a viable and well-documented tectonic mechanism that will allow such underthrusting to occur along a continent margin.

Tectonic implications

The exclusively S-type nature of the West Coast Province batholiths, absence of associated mafic magmatism, derivation of the granitoids by melting of underlying old continental crust as indicated by Nd and Sr isotopes and U-Pb zircon inheritance features, their late Triassic ages and their formation adjacent to the Triassic suture in medial Peninsular Malaysia are best explained by Mitchell's (1977) continent collision model. There is little evidence to support Hamilton's (1979) suggestion that although Triassic granitoids may be present, the batholiths are largely of Jurassic age; that the granitoids represent mixtures of crustal - and mantle - derived magmas and as such, should be related to subduction along a trench system now represented on land by ophiolite - melange rocks in the Shan Plateau of Burma to the northwest.
CHAPTER FIVE

GRANITOIDS BATHOLITHS OF THE EAST COAST PROVINCE

INTRODUCTION

The work summarized by Chappell and White (1974) on the early Palaeozoic batholiths of the Lachlan Fold Belt of southeastern Australia indicates that two contrasting orogenic granitoid types may be recognized. These workers proposed a binary I- and S-type classification of granitoids based on the premise that the mineralogical and chemical characteristics of granitoids are a reflection of their source rocks — primarily, either igneous or sedimentary. Calcic hornblende is a characteristic mineral of I-types but is absent in S-types which are characterized by highly aluminous biotite + cordierite + muscovite + garnet. This conceptual subdivision of orogenic batholiths into two contrasting associations has proved to be very useful despite problems arising from the complexities involved in the generation of I-type magmas and the variable duration and extent of recycling of the sedimentary source rocks for S-type magmas. In this chapter, mineralogical, chemical and isotopic data for the Permian-Triassic batholiths of the East Coast Province of Peninsular Malaysia are presented and used to constrain interpretations concerning the nature of the batholiths, magma sources, granitoid types and granitoid differentiation processes. It will be shown that the basic I-S granitoid subdivision can be recognized even though there are considerable differences between Malaysian and southeastern Australian granitoids of the same type.
Tectonic and Geologic Setting

Aspects of the geology of the Peninsular Malaysia region have been described by Gabbett and Hutchison (1973) and tectonic divisions of the granitoid provinces have been proposed by Hutchison (1977), Mitchell (1977) and Bignell and Snelling (1977). Only a brief summary of the salient aspects of these works are presented here.

The pre-Cenozoic core of southeast Asia has been interpreted to have formed by the late Triassic collision of two continental fragments (Mitchell, 1977). Traces of the collision suture in Peninsular Malaysia is represented by a discontinuous belt of rocks containing elements of the ophiolite-melange association. In his reconstruction, Mitchell (1977) suggested that this ophiolite-melange belt represents the fossil remnants of an Upper Palaeozoic subduction system dipping to the present day east. Permian to Triassic granitoid batholiths and widespread basalt-andesite-dacite-rhyolite volcanics in the East Coast Province would then be the plutonic-volcanic arc complementary to an ophiolite belt in a subduction model.

Granitoid batholiths in the East Coast Province are distributed as linear masses parallel to the medial collision suture in Peninsular Malaysia (Fig. 1). The province extends for a distance of ~600 km and has a typical exposed width of 80 km. The western margin of the province is characterized by an abrupt development of granitoid batholiths along a line approximately 100 km from the medial suture. This 'batholith front' is rather similar to the abrupt volcanic fronts in many subduction settings.
The batholiths are intrusive into generally gently deformed, weakly metamorphosed Carboniferous to Triassic sediments and volcanics. Post-orogenic continental red bed formations of Jurassic-Cretaceous age unconformably overlie both the granitoid batholiths and Palaeozoic rocks.

Although subordinate to the West Coast Province, tin mineralization (together with tungsten and iron mineralization) is important in the East Coast Province and generally occurs in and along the fringes of granitoid outcrops (Rajah et al., 1977).

**Batholith Distribution and Age Patterns**

Informal names of the batholiths shown in Fig.1 are simplified from those of Rajah et al., (1977). The batholiths have not all been mapped in detail and as more work proceeds, better understanding of the internal structure, age patterns, and compositional variations of the constituent plutons may require future changes in the batholith terminology. In common with other terrains, narrow sedimentary screens often separate mappable different batholith units. This argues against any coalescing of the units at depth. The batholiths are surrounded by narrow contact metamorphic aureoles superimposed on low-grade metamorphic country rocks (Hutchison, 1977).

A biotite + hornblende granodiorite-adamellite of variable texture is the most common rock type but a single intrusive complex may consist of rocks ranging from gabbro to potassic granite. Where found, mafic plutons are the earliest in a sequence of intrusions and they occur as marginal bodies intruded by the widespread felsic plutons. Mafic dyke swarms are common.
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K-Ar cooling ages are generally concordant with more limited Rb-Sr whole rock isochron and U-Pb zircon ages (Bignell and Snelling, 1977; chapter three). The radiometric dating results summarized in Fig. 2 indicate ages of emplacement of the batholiths from Permian to Triassic.* This is consistent with the known stratigraphic relationships. Radiometric ages display a general younging trend towards the medial suture in Peninsular Malaysia and is indicative of a westward migration of plutonic locus with time. Three intrusive episodes are suggested by 'breaks' in the present K-Ar data (Fig. 2). These suggested episodes are: 255-270 Ma (early Permian) and 220-240 Ma (late Permian-early Triassic) in the East Coast Province and 200-220 Ma (middle-late Triassic) for the adjacent Central Province. The most important episode was the late Permian-early Triassic episode and intrusions of this age range are found along the entire north-south length of the 600 km batholith terrain.

**Samples and Experimental Techniques**

Fifty-four samples representative of rock types ranging from gabbro to granite were collected for thin-section examination, microprobe analyses and comprehensive chemical analyses. Mineral composition data reported here are obtained with an Si (Li) detector-equipped energy dispersive electron microprobe at the A.N.U. Analytical procedures, precision and accuracy of the system are discussed by Reed and Ware (1975).

* For consistency, the time scale adopted in this work follows that of Bignell and Snelling (1977) but their ages have been recalculated using the new decay constants recommended by Steiger and Jager (1977). The Triassic/Jurassic boundary is taken at 195 Ma, Triassic/Permian boundary at 250 Ma and the Carboniferous/Permian boundary at 280 Ma.
Fig. 2. Summary of the available geochronological data for the East Coast and Central Province granitoids plotted to illustrate the variation of ages with distance from the medial suture. K-Ar and much of the Rb-Sr data are from Bignell and Snelling (1977). U-Pb zircon results were discussed in chapter three. Symbols are joined by short lines where age determinations were made on a single pluton or a cogenetic suite using different radiometric methods. Interpreted intrusive ages display a general younging trend westwards. East-west distance is approximately 150 km.
H₂O and CO₂, the chemical data were obtained by X-ray fluorescence spectrometry. Major elements were determined on fused glass discs and trace elements on pressed powder pellets. Details of procedures, instrumental conditions and correction techniques are discussed in Norrish and Chappell (1977). In addition, selected samples were analysed for REE by instrumental neutron activation procedures. Precision and accuracy of the technique are estimated to be better than 10% for the REE.

Representative analyses of S- and I-type plutons from the East Coast Province are given in Table 1 and their sample localities are listed in the appendix.

GENERAL MAJOR ELEMENT CHEMISTRY

On a conventional AFM diagram (Fig.3), the East Coast Province batholiths display a deceptively simple calcalkaline trend. However a frequency plot (Fig.4), using the present data together with representative data from previous published studies, reveals a bimodal distribution of silica values and the scarcity of intermediate compositions with 55-65% SiO₂. Because mafic rocks (<55% SiO₂) make up <5% of the area of the exposed batholiths, any suggestion for an origin of the widespread I-type granodiorite plutons based on differentiation (+ crustal assimilation) from mafic or intermediate parents is thus confronted with the problem of having to explain the scarcity of possible parent compositions and why inferred fractionated compositions form a separate frequency mode.
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Fig. 3. Analysed samples from the East Coast Province plotted on an AFM diagram.
Fig. 4. Histogram plot of SiO₂ contents of granitoids and mafic rocks of the East Coast Province. Incorporates representative data from Hutchison (1973), Rajah et al. (1977) and Kumar (1980, 1981).
On a modified ACF diagram (Fig. 5), the analyses plot predominantly in the biotite-plagioclase-hornblende triangle but some plot above the biotite-plagioclase join indicating that they are peraluminous. Based on other criteria discussed later, two granitoid fields can be outlined—one defined by the dominant metaluminous I-type suites which may fractionate to peraluminous granitic compositions and a second field for peraluminous granodiorite–granite S-type suites.

Fig. 6 compares CaO versus SiO₂ and K₂O versus SiO₂ plots of the East Coast plutonic province with those of other important Circum-Pacific batholith terrains. These plots show that the East Coast Province batholiths are among the most potassic and least calcic of the terrains compared. The highly siliceous, high K₂O and low CaO overall nature of the province suggests that the batholiths may have been formed in a region overlying thick underlying continental crust. This suggestion is also supported by Nd-Sr isotopic and U-Pb zircon data (chapter six).

MINERAL CHEMISTRY

I-type granitoids in the East Coast Province characteristically contain biotite ± hornblende. Hornblende is absent in S-types but distinctive S-type mineral indicators like primary muscovite, garnet, cordierite and sillimanite are absent. Mineral compositional arguments used to distinguish I- and S-type granitoids in the East Coast Province are presented in this section.

I-type granitoids may occur as isolated plutons of variable size, as the dominant components of complexes made up of nested units of gabbroic rocks and granodiorite–granite plutons and as large plutons
Fig. 5. Analysed samples from the East Coast Province plotted on a modified ACF diagram (White and Chappell, 1977). Symbols as for Fig. 3.
Fig. 6A-B. Granitoids of the East Coast Province compared with generalized fields for granitoids of the Uasilau-Yau Yau area of New Britain (Whalen, 1980), Berridale-Kosciusko batholiths of the Lachlan Fold Belt of eastern Australia (Hine et al., 1978) and Sierra Nevada of western U.S.A. (Bateman and Dodge, 1970) on CaO vs. SiO$_2$ and K$_2$O vs. SiO$_2$ variation diagrams. Symbols as for Fig. 3.
Table 2: Representative microprobe analyses of amphiboles

|         | 93   | 94 | 95   | 99 | 100 | 101 | 103 | 104 | 105 | 106 | 107 | 108 | 109 | 110 | 111 | 112 | 113 | 114 |
|---------|------|----|------|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| SiO₂    | 47.90| 39.88| 48.39| 44.62| 46.90| 41.75| 48.45| 46.45| 41.94| 47.82|
| TiO₂    | 0.69 | 1.35 | 0.65 | 1.33 | 0.92 | 1.95 | 1.03 | 1.87 | 0.89 |
| Al₂O₃   | 6.47 | 8.71 | 5.24 | 8.35 | 6.13 | 8.22 | 6.18 | 8.79 | 5.81 |
| FeO     | 16.85| 28.41| 14.82| 21.81| 16.89| 25.22| 18.96| 25.49| 17.69|
| MnO     | 0.57 | 1.16 | 0.56 | 0.53 | 0.82 | 0.71 | 0.41 | 0.77 | 0.50 |
| MgO     | 12.47| 3.17 | 13.95| 8.88 | 11.80| 4.83 | 11.06| 5.06 | 12.22|
| CaO     | 11.29| 10.62| 12.05| 10.16| 11.48| 10.26| 11.08| 10.47| 10.80|
| K₂O     | 0.58 | 1.48 | 0.46 | 0.49 | 0.68 | 1.20 | 0.65 | 1.30 | 0.50 |
| Na₂O   | 1.05 | 2.05 | 0.76 | 1.27 | 1.19 | 1.76 | 1.18 | 1.81 | 1.21 |
| Cl      | 0.46 | 0.13 | -    | -    | 0.10 | 0.65 | 0.26 | 0.10 | 0.21 |

*total Fe as FeO

Table 3: Representative microprobe analyses of biotites

|         | 93   | 94 | 95   | 99 | 100 | 101 | 103 | 104 | 105 | 106 | 107 | 108 | 109 | 110 | 111 | 112 | 113 | 114 |
|---------|------|----|------|----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
| SiO₂    | 36.17| 33.70| 36.53| 35.13| 36.55| 35.02| 35.57| 34.57| 35.82| 34.04| 34.26| 34.64| 34.47|
| TiO₂    | 4.18 | 3.01 | 4.46 | 2.98 | 3.71 | 3.44 | 3.48 | 3.76 | 3.63 | 3.03 | 2.43 | 3.31 |
| FeO*    | 20.36| 34.02| 18.38| 25.32| 20.20| 27.60| 27.71| 23.29| 28.62| 24.92| 29.45| 23.44|
| MnO     | 0.24 | 0.85 | 0.27 | 0.27 | 0.57 | 0.35 | 0.31 | 0.18 | 0.42 | 0.16 | 1.32 | 0.44 |
| MgO     | 11.06| 1.58 | 12.47| 7.41 | 11.33| 5.28 | 5.55 | 10.31| 2.77 | 5.72 | 2.27 | 5.36 |
| Na₂O   | 0.21 | 0.23 | 0.17 | 0.26 | 0.21 | 0.18 | 0.20 | 0.16 | 0.24 | 0.28 | 0.12 | 0.17 |
| Cl      | 0.36 | 0.27 | 0.10 | 0.12 | 0.49 | 0.10 | 0.52 | 0.24 | 0.28 | 0.12 | 0.17 |

Table 4: Representative microprobe analyses of Fe-Ti oxides

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<th>ilmenites(S-types)</th>
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in complex composite batholiths that may also contain gabbros and S-type plutons. Mineralogical assemblages are invariably some permutation of the theme plagioclase + K-feldspar + quartz + hornblende + biotite + magnetite + ilmenite. Iron sulphide, zircon, apatite, sphene, and allanite are typical accessory minerals. Sphene can achieve the status of a major mineral in some plutons.

S-type granitoids occur as scattered stocks and as constituent plutons of large batholiths. Mineralogy is characterised by plagioclase + K-feldspar + quartz + high-Al biotite + ilmenite + tourmaline. Subsolidus muscovite may be present. Iron sulphides, zircon and apatite are the common accessory phases.

Feldspars

Plagioclase grain relationships indicate extended plagioclase crystallization histories in both S- and I-type plutons. Calcic plagioclase laths observed as inclusions in early euhedral amphibole phenocrysts and clots of tabular calcic plagioclase phenocrysts suggest that plagioclase was the earliest crystallizing phase in the I-type granitoid magmas. Late crystallization of plagioclase is evident in porphyritic granites where sodic plagioclase + K-feldspar + quartz make up the equigranular groundmass. Normal and oscillatory zoning in plagioclase is ubiquitous. Also present is a stepped zoning pattern where large breaks in anorthite content occur between adjacent zones. Grains with highly irregular, mottled and inclusion-charged cores are present in many plutons and such grains have been interpreted to represent the incorporation of restitic (source residue after partial melting) plagioclase (Presnall and Bateman, 1973; White and Chappell, 1977).
The most calcic plagioclase analysed is An\textsubscript{65} in an I-type quartz monzodiorite. The most calcic plagioclase analysed in an S-type granitoid is An\textsubscript{42}. A plot of mol. %An range in plagioclase against whole rock % Si\textsubscript{2}O\textsubscript{2} is shown in Fig. 7 which can be used as a crude guide to the variation of plagioclase compositions with rock type. The typical range of mol. %An of plagioclase in a single thin-section estimated by core and rim compositions is 10 - 20%. Oligoclase-andesine, An\textsubscript{20} - An\textsubscript{40}, is the typical plagioclase in the widespread granodiorite plutons.

The variability and complexity of the plagioclase features make any generalization difficult. Observed features can be explained by any combination involving early crystal accumulation, reaction with changing melt compositions, incorporation of source plagioclase and even magma mixing.

Alkali feldspars have very high Or contents and are typically Or\textsubscript{80}Ab\textsubscript{20} to Or\textsubscript{90}Ab\textsubscript{10}. Extensive exsolution from more Ab-rich high temperature forms is indicated. The late interstitial habit of K-feldspar in many I-type plutons contrast strongly with the megacrystic habit in some S-type plutons.

**Amphiboles**

Amphiboles are restricted to I-type plutons. They may occur as euhedral grains containing inclusions of calcic plagioclase and opaque oxides or as ragged grains associated with biotite and magnetite in mafic clots. Such clots, together with less common enclaves of granoblastic plagioclase + hornblende + biotite may plausibly be interpreted as modified restitic assemblages. Pyroxene cores in amphiboles are extremely rare and there is little evidence to support a reaction relationship between liquid and pyroxene to form amphibole
Fig. 7. Plot of %An (plagioclase) vs. whole rock SiO₂ content for some East Coast granitoids. Vertical lines represent ranges in An contents estimated using core and rim microprobe determinations. Solid circles = S-types, empty circles = I-types. Refer to appendix 7 for numbered sample localities.
in the vast majority of the plutons. Irregular overgrowths of late amphibole over early euhedral amphibole prisms is a texture recording breaks in crystallization histories. Very late, Fe-rich interstitial amphibole is present in highly siliceous granitic plutons.

Analysed amphiboles are calcic (atomic Ca>1.34 per 23 oxygens), have \((Na + K)_A\) generally < 0.50 per 23 oxygens and atomic Ti < 0.50 per 23 oxygens. Following Leake's (1978) classification (Fig.8), they may be termed magnesio-hornblendes and ferro-hornblendes but these grade into amphiboles with \((Na + K)_A > 0.5\) per 23 oxygens in more siliceous plutons. These latter amphiboles plot in the fields of ferro-edenite and ferro-edenitic hornblendes. A plot of Mg/Mg+Fe (amphibole) against whole rock % SiO₂ (Fig.9) shows a clear iron-enrichment trend. Hornblendes from plutons containing early sphene + magnetite are considerably more magnesian than other I-type plutons with comparable whole rock SiO₂ values and plot significantly above the field of hornblendes from sphene-absent plutons which contain magnetite or ilmenite or both. The high Mg-hornblendes thus corroborate the intrinsically more oxidized nature of plutons containing early sphene and magnetite (Czamanske et al., 1981). Hornblendes from the sphene-absent plutons plot near the limited trend shown by the Takanawa Peninsula ilmenite (reduced)-series granitoid hornblendes studied by Czamanske et al., (1981). But unlike the Japanese examples, the hornblendes studied here reach extremely Fe-rich compositions \(\text{Mg}/(\text{Mg} + \text{Fe}) = 0.15\) for sample 94). Such Fe-enrichment trends indicate that the studied plutons did not develop increasingly oxidized conditions (e.g. by differential loss of hydrogen) with fractionation as this would yield increasing or constant \(\text{Mg}/(\text{Mg} + \text{Fe})\) ratios in hornblende with increasing whole rock % SiO₂ as Czamanske et al., (1981) argued.
Fig. 8. Microprobe analyses of amphiboles plotted on Leake's (1978) classification diagram.

Fig. 9. Plot of atomic Mg/(Mg+Fe) of amphibole vs. whole rock SiO₂ content for representative granitoid samples. Numbered localities listed in appendix 7.
Euhedral hornblende prisms with laths of plagioclase inclusions are textural evidence that indicate crystallization of hornblende from a melt. For hornblendes to crystallize in felsic magmas, Burnham (1979) argued that the water content of the melt phase must exceed \( \sim 3 \) wt. %. Early crystallization of plagioclase before hornblende observed in thin section can potentially be used to constrain the upper limits of initial water contents of the East Coast Province magmas. For this order of crystallization i.e. plagioclase before hornblende, the experimental results of Naney and Swanson (1980) suggest \( H_2O < 5 \) wt. % for a granodiorite at 8 kbar and the work of Wyllie (1977) for a tonalite at 10 kbar suggests \( H_2O < 3 \) wt. %.

These results indicate that I-type magmas of the East Coast Province may have had original water contents of \( \sim 3 \) wt. % in the melt phase. Additional solidification would increase this value sufficiently to stabilize early hornblende after early plagioclase.

**Biotites**

Biotites occur as discrete plates of variable size, as ragged grains in composite mafic clots, as inclusions in quartz, K-feldspar and more rarely, plagioclase and occasionally as rims around amphiboles (in I-type plutons).

Analysed biotites are characterized by high \( K_2O \) (> 9%) and high \( TiO_2 \) (3.0 - 4.5%). On a plot of \( Fe/(Fe+Mg) \) against \( Al^{4} \) (Fig.10), biotite compositions can be observed to be more aluminous than that defined by the phlogopite - annite join. Biotites in felsic plutons trend towards the siderophyllite corner. A plot of \( Fe/(Fe+Mg) \) against \( Al^{6} \) (Fig.11) provides ideal separation of biotites from I- and S-type plutons. Biotites from I-type plutons have \( Al^{6} \) varying from 0.01 to 0.25 atoms per 22 oxygens whereas the Fe-rich aluminous
Fig. 10. Plot of atomic Al$^4$ vs. Fe/(Fe+Mg) of biotites from the East Coast Province. Solid circles = S-types, empty circles = I-types.

Fig. 11. Plot of atomic Al$^6$ vs. Fe/(Fe+Mg) of biotites. Symbols as for Fig. 10.
Biotes of S-type plutons possess Al\textsuperscript{6} contents ranging from 0.50 to 0.80 atoms per 22 oxygens. Biotites from I-type plutons show increasing Al\textsuperscript{6} with increasing differentiation whereas biotites from S-type plutons show decreasing Al\textsuperscript{6} with increasing differentiation (using whole rock % SiO\textsubscript{2} as differentiation index). The distinct biotite compositional fields argue strongly against a mechanism for producing the volumetrically minor S-type plutons of the East Coast Province from I-type parents by large amounts of assimilation of metasedimentary country rocks.

The iron-enrichment trend with increasing whole rock % SiO\textsubscript{2} for biotites parallels that described for hornblendes (Fig.12). For comparable SiO\textsubscript{2} contents, biotites from sphene + magnetite - bearing I-type plutons are characterized by significantly higher magnesium contents compared to biotites from other plutons.

**Opaque mineralogy**

Ilmenite, magnetite, pyrite and pyrrhotite are the opaque minerals present. Opaque assemblages in I-type plutons are ilmenite + Fe-sulphide, ilmenite + magnetite + Fe-sulphide and magnetite + Fe sulphide. Ilmenite + Fe-sulphide is the opaque mineral assemblage in the S-type plutons. Pyrrhotite is the more common Fe-sulphide phase in the East Coast Province plutons.

The most common textural habit of magnetite is as prominent euhedral-subhedral aggregates or discrete grains associated with amphibole and biotite. Less commonly, magnetite is observed in quartz and feldspar. Ilmenite grains often occur as isolated rods and anhedral blebs included in mafic silicates. These textural habits of magnetite and ilmenite are interpreted as indicating a magmatic origin
Fig. 12. Variation of atomic Mg/(Mg+Fe) of biotite with whole rock SiO₂ content of representative samples. Symbols as in Fig. 10.

Fig. 13. Variation of atomic Mn in coexisting biotite and ilmenite for the East Coast Province granitoids and other granitoid terrains where such data were available.
and reported analyses are restricted to such grains. Dense aggregates of small grains associated with alteration of the mafic silicates were avoided.

Magnetites analysed have low TiO₂ (< 1 wt. %) and MnO (< 0.8 wt. %) contents. The low TiO₂ contents invalidate the use of the magnetite - ilmenite geothermometer - oxybarometer of Buddington and Lindsley (1964) and indicate that the magnetites have equilibrated to low temperatures. Calculated mol. % 'Usp' \( \frac{\text{Fe}_2\text{TiO}_4}{(\text{Fe}_2\text{TiO}_4 + \text{Fe}_3\text{O}_4)} \) range from < 1 to 3. Magnetite seems to be restricted to I-type plutons.

Ilmenites from both S- and I-type plutons contain a significant MnO component (4 – 13 wt. %). Calculated values of mol. % 'Ilm' \( \frac{\text{FeTiO}_3}{(\text{FeTiO}_3 + \text{Fe}_2\text{O}_3)} \) range from 90 to 99 for four I-type plutons and 98 to 100 for three S-type plutons. Some workers (e.g. Czamanske et al., 1981) have suggested that bulk ilmenite compositions of plutonic rocks may still retain a memory of magmatic oxygen fugacities even though coexisting magnetites indicate subsolidus re-equilibration to very low temperatures. Several observations suggest that this may be a reasonable assumption. Firstly, a plot of Mn contents of coexisting ilmenite and biotite (Fig.13) for the East Coast Province plutons show that they fall within a correlation field suggestive of magmatic equilibration between coexisting ilmenite and biotite pairs from southeastern Australia (Whalen, 1980), Japan (Czamanske et al., 1981) and the West Coast Province of Peninsular Malaysia (chapter four). Secondly, there is a systematic and consistent increase in Fe₂O₃ contents of ilmenites from plutons lacking magnetite including S-types (116 : Ilm₁₀₀; 133 : Ilm₉₈; 134 : Ilm₉₉) and I-types
(103: Ilm_{96-97}; 125: Ilm_{99}; 141: Ilm_{94}) to a magnetite-bearing I-type (95: Ilm_{90-91}). Fig.14 shows several representative stability curves of the ilmenite compositions obtained together with some relevant buffer curves on an fO₂ - T(°C) plot. Taken at face value, the ilmenite results suggest that for reasonable temperatures of 700 - 900°C, I-type plutons have experienced and may have crystallized under very variable fO₂ conditions ranging over four orders of magnitude. A minimum estimate for the highly oxidized sphene + magnetite-bearing plutons is represented by the hedenbergite + ilmenite= quartz + magnetite + sphene reaction curve. This fO₂ - T(°C) curve drawn in Fig.14 is that calibrated by Wones (in prep.). Ilmenites from S-type plutons consistently indicate low fO₂ conditions as evident by in their Ilm_{98-100} compositions. The stability curves of such ilmenite compositions would lie near the magnetite-wustite and graphite - CO₂ buffer curves.

It is concluded that, if the assumption of inert behaviour of Mn is valid and the ilmenites are recording magmatic conditions, then S-type plutons have crystallized under reduced conditions near the graphite - CO₂ buffer probably because of their derivation from graphite-bearing metasedimentary sources. I-type plutons have crystallized under variable fO₂ conditions ranging from reduced conditions near magnetite-wustite to oxidizing conditions in the vicinity of the sphene-magnetite fO₂ - T(°C) curve. The absence of evidence for late-stage oxidation as indicated by consistent mafic silicate iron-enrichment trends and the Fe³⁺/total Fe versus SiO₂ variation diagram (Fig.15) suggests that the fO₂ conditions obtained from ilmenite data for the I-type plutons may represent those of the source regions.
Fig. 14. Probable fields for I-type and S-type granitoids of the East Coast Province on an $f_{O_2}$ - $T$ (°C) plot based on ilmenite composition data (see text and table 2). Sphene + magnetite curve is after Wones (in prep.). Ilmenite stability curves are those of Spencer and Lindsley (1981). See Ohmoto and Kerrick (1977) for references to other buffer curves.
Fig. 15. Plot of atomic $\text{Fe}^{3+}/\text{total Fe}$ vs. $\text{SiO}_2$ for granitoids of the East Coast Province.
MAJOR AND TRACE ELEMENT GEOCHEMISTRY

The major and trace element characteristics of the I- and S-type granitoids distinguished based on petrography and mineral composition data are presented in this section. The chemical data are then used to place constraints on granitoid differentiation processes.

Selected major and trace elements plotted against SiO$_2$ on conventional Harker variation diagrams are shown in Fig.16. For the I-type association, K$_2$O increases with increasing SiO$_2$, Na$_2$O remains relatively constant whereas CaO, MgO, FeO, Al$_2$O$_3$, P$_2$O$_5$ and TiO$_2$ all decrease with increasing SiO$_2$. The behaviour of the trace elements plotted on Harker variation diagrams can be divided into three groups:

1. incompatible trace elements that increase with increasing SiO$_2$: Pb, Rb, Ba
2. compatible trace elements that decrease with increasing SiO$_2$: Zr, V, Ni, Cr, Zn, Sr
3. decoupled trace elements that show no correlation with SiO$_2$: Sn, Nb, U, Y, Cu, Ga, Th, REE

Major and trace element trends of S-type granitoids are similar to I-type trends with the exception of Ba and LREE which decrease with increasing SiO$_2$ and U, Th which increase with increasing SiO$_2$. For comparable SiO$_2$ contents, S-types have higher abundances of Th, U, Pb and Y than I-types.
Fig. 16 See following page
Fig. 16. Harker variation diagrams for selected major and trace elements. Markings on vertical axes for trace elements in ppm, for major element oxides in Wt. %. Fields outlined in Ba vs. SiO₂ plot illustrates contrasting trends of several I-type suites.
The behaviour of Ba is particularly enlightening and highlights an important difference in the mode of fractionation between S- and I-types suites. In S-type suites and in one I-type pluton, Ba decrease with increasing SiO₂, in some samples, down to levels as low as 50 ppm. In contrast, I-type suites show an increase from ~600 - 800 ppm Ba at 68% SiO₂ to 1000 - 1200 ppm at 76% SiO₂ (Fig.14). Biotite is the only major mineral phase capable of explaining this phenomena. Because of the large value of the partition coefficient for Ba of biotite, Ba is preferentially partitioned into biotite relative to the melt. Thus, crystallization of early biotite in S-type magmas depletes the residual melt in Ba whereas late crystallization of biotite in I-type magmas causes enrichment of early and intermediate derivative melts in Ba.

The erratic behaviour of the LREE has been suitably explained by Miller and Mittlefehldt (1982). They suggested that the variable removal of accessory phases like sphene, allanite, apatite and monazite which contain the bulk of the LREE will determine whether the LREE increase, decrease or are decoupled with increasing fractionation.

Other trends like the compatible behaviour of Y, Ni, Cr, Zn and decreasing MgO, FeO and TiO₂ with increasing SiO₂ can be readily explained by the removal of hornblende, biotite and opaque oxides. The negative trends of CaO, Sr with increasing SiO₂ is attributed to plagioclase fractionation whereas zircon and apatite removal can explain the negative trends of Zr and P₂O₅ respectively.

In principle, then, the behaviour of the major and trace elements can be qualitatively explained by hornblende - biotite - plagioclase - Fe-Ti oxide fractionation for the I-type granitoids and biotite -
plagioclase - ilmenite fractionation for the S-type granitoids. Zircon and apatite are accessory phases removed by fractionation in both granitoid types. Variable removal or incorporation of allanite and sphene for I-type granitoids and monazite for S-type granitoids can explain the variable behaviour of LREE, U and Th.

Crystal fractionation models

Two theoretical end member crystallization modes that can be considered are perfect equilibrium crystallization for the case when the solid phase remains at all time in equilibrium with the melt and perfect (Rayleigh) fractional crystallization for the case when the solid phase is prevented from reacting with the melt once it is formed. Previous geochemical fractionation models for granitoids (e.g. McCarthy and Hasty 1976), the ubiquity of zoned plagioclases, the absence of cumulates associated with granite plutons and the viscous nature of semi-liquid granitoid magmas that makes crystal accumulation prohibitive all argue for an intermediate crystallization mode.

Such an intermediate crystallization mode can be considered as an incremental equilibrium crystallization process and its behaviour is described by the equation:

\[
\frac{C_n}{C_0} = \prod_{i=1}^{n} \left[ \frac{1}{f_i (D_i - 1) + 1} \right]
\]

where \( C_n \) = concentration of a trace element in the \( n \)th increment liquid; \( C_0 \) = original concentration of the element in the liquid; \( f_i \) = fraction of solid removed relative to the mass of the \((i-1)\)th liquid; \( D_i \) = bulk solid/liquid partition coefficient during the \( i \)th increment, calculated from \( D = \sum W_i K_i \) where \( W_i \) is the weight of
a mineral phase and \( K_d \) is the mineral/liquid partition coefficient. The concentration of a trace element in the \( i^{th} \) solid increment is given by:

\[
C_i^S = D_i \cdot C_i^L
\]  

Equation (1) represents a general case of crystallization behaviour which at one limit, a very large number of infinitely small increments describe perfect fractional crystallization and at the other limit, a single increment describes perfect equilibrium crystallization. By choosing appropriate values of \( f_i \), a variety of incremental equilibrium crystallization behaviour may be described.

For the modelling of crystallization behaviour of the East Coast Province granitoid magmas, Rb, Ba and Sr were the elements selected because they are compatible trace elements with respect to one or more of the major silicate minerals plagioclase, biotite and K-feldspar. They do not form accessory phases and are not important in the typical accessory minerals whose modal abundances are so difficult to estimate. They can, therefore, be used as more reliable indicators of large-scale directions of magmatic changes compared to elements like the REE, U and Th. Partition coefficients used in this study are listed in table 6. Variations in actual coefficients are well known but these variations will not affect the main conclusions as we are discussing chemical trends rather than rigorous quantitative details.

(A) I-type fractionation processes

Several unknowns are involved in the computation of closed system fractionation models. These are the proportions of various crystallizing phases, the incoming of new crystallization phases and the composition of the initial liquid. To constrain these unknowns, I
have adopted the approach of using major element modelling results to yield qualitative estimates of the proportions of various phases fractionated and assumed that hornblende, plagioclase, biotite and quartz are the major minerals controlling fractionation trends. The assumption that pyroxene and olivine are not involved is reasonable in view of the bimodal separation of gabbroic and granitoid associations in the East Coast Province and the minor volumes of mafic and intermediate plutons. The method of computation for major element modelling essentially involves subtraction calculations of the major element oxide compositions of crystallizing minerals from a parental liquid composition. By varying the proportions of the various crystallizing phases, a qualitative fit to the observed oxide trends can be generated. The only constraint is the balancing of the relationship:

\[
\% \text{oxide in parental liquid} = \% \text{oxide of bulk solid assemblage} + \% \text{oxide in derivative liquid}
\]

Compositions of subtracted mineral phases vary with degree of fractionation and actual compositions used are those discussed in the previous section on mineral chemistry.

Results of a typical fit to the actual major element oxide trends starting with a hypothetical 65% $SiO_2$ (average of the most mafic granitoid plutons) parent are shown in tables 5A-B together with the mineral assemblages subtracted after 10% solidification intervals of a derivative liquid. Several significant comments can be made based on the major element oxide modelling at this stage.

1. It is stressed that we are not modelling the evolution of a single zoned pluton (e.g. Loch Doon pluton by Tindle and Pearce, 1981). The solution is not unique and its use is primarily that of a
### Table 5a: Fractionating phases utilised for major element oxide and trace element modelling of I-type trends

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<td>0.10</td>
<td>0.10</td>
<td>0.05</td>
</tr>
<tr>
<td>K-feldspar</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>7.5 x 10^-3</td>
<td>7.5 x 10^-3</td>
<td>5 x 10^-3</td>
<td>5 x 10^-3</td>
<td>5 x 10^-3</td>
</tr>
<tr>
<td>Magnetite</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>7.5 x 10^-3</td>
<td>7.5 x 10^-3</td>
<td>5 x 10^-3</td>
<td>5 x 10^-3</td>
<td>5 x 10^-3</td>
</tr>
<tr>
<td>Ilmenite</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>7.5 x 10^-3</td>
<td>7.5 x 10^-3</td>
<td>5 x 10^-3</td>
<td>5 x 10^-3</td>
<td>5 x 10^-3</td>
</tr>
</tbody>
</table>

* Each step represents removal of 10% wt. fraction from preceding, not original, liquid as crystals. Mass of original liquid remaining = 0.9^i where i = i**th** removal step.

### Table 5b: Typical evolving liquid compositions obtained from major element oxide modelling

<table>
<thead>
<tr>
<th>0</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
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</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>65.0</td>
<td>66.3</td>
<td>67.6</td>
<td>69.4</td>
<td>70.6</td>
<td>71.8</td>
<td>73.0</td>
<td>74.0</td>
<td>75.0</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.2</td>
<td>15.6</td>
<td>15.1</td>
<td>14.5</td>
<td>14.1</td>
<td>13.7</td>
<td>13.2</td>
<td>12.8</td>
<td>12.5</td>
</tr>
<tr>
<td>FeO</td>
<td>3.8</td>
<td>3.5</td>
<td>3.1</td>
<td>2.7</td>
<td>2.4</td>
<td>1.9</td>
<td>1.4</td>
<td>0.85</td>
<td>0.60</td>
</tr>
<tr>
<td>MgO</td>
<td>1.8</td>
<td>1.5</td>
<td>1.2</td>
<td>1.0</td>
<td>0.8</td>
<td>0.6</td>
<td>0.4</td>
<td>0.2</td>
<td>0.1</td>
</tr>
<tr>
<td>CaO</td>
<td>4.8</td>
<td>4.1</td>
<td>3.4</td>
<td>2.8</td>
<td>2.3</td>
<td>1.9</td>
<td>1.6</td>
<td>1.4</td>
<td>1.1</td>
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<tr>
<td>K₂O</td>
<td>2.5</td>
<td>2.8</td>
<td>3.1</td>
<td>3.4</td>
<td>3.7</td>
<td>4.0</td>
<td>4.3</td>
<td>4.5</td>
<td>4.7</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.5</td>
<td>3.45</td>
<td>3.40</td>
<td>3.26</td>
<td>3.11</td>
<td>3.0</td>
<td>2.9</td>
<td>2.7</td>
<td>2.5</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.5</td>
<td>0.48</td>
<td>0.46</td>
<td>0.43</td>
<td>0.40</td>
<td>0.35</td>
<td>0.30</td>
<td>0.25</td>
<td>0.20</td>
</tr>
</tbody>
</table>

### Table 6: Mineral partition coefficients used for LIL element modelling

<table>
<thead>
<tr>
<th></th>
<th>Plagioclase</th>
<th>Biotite</th>
<th>K-feldspar</th>
<th>hornblende</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>dacite</td>
<td>rhyolite</td>
<td>dacite</td>
<td>rhyolite</td>
</tr>
<tr>
<td>Rb</td>
<td>0.05</td>
<td>0.02</td>
<td>3.3</td>
<td>2.0</td>
</tr>
<tr>
<td>Ba</td>
<td>0.4</td>
<td>1.3</td>
<td>6.4</td>
<td>10</td>
</tr>
<tr>
<td>Sr</td>
<td>3</td>
<td>5</td>
<td>0.12</td>
<td>3.9</td>
</tr>
</tbody>
</table>

From compilations of Cox et al. (1979), Perfit et al. (1980) and Ewart (1982).
qualitative guide to the realistic proportions of mineral phases fractionated during the evolution of tonalitic-granodioritic liquids to granitic liquids.

2. Major element oxide trends can be modelled by fractionating plagioclase (in proportions decreasing from 70% to 50%), amphibole (decreasing from 30% to 5%), biotite (increasing from 5% to 15%), quartz (increasing from 5% to 20%) and late K-feldspar (0% to 10%). The plagioclase-amphibole proportions are approximately similar to those discussed by Perfit et al., (1980a) in their models to generate dacitic liquids from andesitic liquids. Quartz is required in the present model to obtain an acceptable fit to observed oxide trends.

3. The results shown in table 5B and illustrated in Fig.17 indicate that for 65 - 70% SiO₂ parental melts, excessive amounts (between 40 to 60% of the original mass) of granitic liquids (> 75% SiO₂) would be generated in cumulate-type models involving the production of cumulate sequences at depth and emplacement of largely liquid magmas. Resorting to more mafic parental liquids (i.e. <65% SiO₂) would produce the intermediate magmas which are not observed in this plutonic province. Purely cumulate models are thus not satisfactory and generally not applicable to granitoid suites. The general consensus (Presnall and Bateman, 1973; White and Chappell, 1977; Pitcher, 1979; Hildreth, 1981 and many others) is that granitoid magmas evolve by incomplete and variable separation of liquid, liquidus crystals and refractory crystalline residue (or restite). The operation of such a process in high-level magma chambers can explain the rock type abundances of the East Coast Province. In particular, the differentiation of
Fig. 17. Diagram illustrating variations in SiO$_2$ contents of derivative liquids from a 65% SiO$_2$ (curve A) parent liquid and 69.5% SiO$_2$ (curve B) parent liquid based on simple fractional crystallization model calculations summarized in table 5 A-B. Lengths of vertical lines at 77% SiO$_2$ correspond to mass fractions of original liquid remaining that have SiO$_2$ contents > 77% SiO$_2$. Dashed curve B' represents the case when variable amounts of liquidus crystals and restite are retained in the liquids represented by curve B. Mass fractions of incorporated solids (assumed to have constant SiO$_2$ contents of 55%) are shown beside curve B'.
an original 70% SiO₂ liquid with variable incorporation (10 - 30%) of crystallized solids and restitic assemblages modelled as curve B' in Fig.17 predicts, for in situ crystallization models (i.e. negligible hidden cumulate bodies at depth), ~ 40% 'granodioritic' (65 - 70% SiO₂) magmas, ~ 40% 'adamellitic' (70 - 75% SiO₂) magmas and ~ 20% 'granitic' (> 75% SiO₂) magmas. The dominance of granodiorite-adamellite compositions predicted by this general model is compatible with the observation that granodiorites and adamellites are the most widespread rock types in the East Coast Province (Fig.4).

4. Fig.17 also shows that highly siliceous granitic liquids (75 - 77% SiO₂) can be obtained after only 40 - 60% of an original 70% SiO₂ liquid had crystallized.

Figs.18 A-C display the interelement Rb - Ba - Sr variations and results of the trace element modelling using equations (1) and (2). The representative I-type fractionation curves shown in each plot are those constrained by major element modelling results. These are explained in more detail in the figure captions. Abundances of Rb, Ba and Sr for ~ 70% SiO₂ samples were first selected for trial computations as estimates of primitive liquid abundances and then refined to fit the actual LIL element trends observed. Only values of f (fraction of initial liquid solidified) from 0.1 to 0.6 are shown because firstly, it became clear during the computations that correspondence to actual trends as well as abundances are obtained after f = 0.6. Secondly, major element oxide models show that the most felsic granites can be obtained after f = 0.4 - 0.6 and furthermore, that incomplete unmixing of restite and crystallized
Fig. 18 A-C. Rb-Ba-Sr variation plots for the East Coast Province granitoids. Ruled triangular areas in each figure represent model starting liquid compositions. For I-type granitoids, removal of plagioclase, hornblende, biotite and quartz causes liquids to evolve along fractionation curves similar to those numbered 1-4. Fractionation curve 3 represents the case where biotite crystallization is initiated very late – this causes rapid enrichment in Ba for fractionated liquids. Curve 4 represents the case where hornblende is the only fractionated hydrous silicate phase. Ticks along fractionation curves denote 10% solidification intervals of residual liquids. In terms of original liquid remaining, the ticks correspond to 0.9, 0.81, 0.73, 0.66, 0.59, 0.53, 0.48, 0.43 and 0.39 mass fraction intervals. Dashed lines enclose areas representing fractionated solid compositions. Variable retention of solids in fractionating liquids yield points that plot between liquid fractionation curves and solid fractionation curves. Highly fractionated I-types (75-77% SiO₂) are enclosed in a dotted field. Note that a handful of high-Sr I-types cannot be explained by the general fractionation model discussed. Symbols as in Fig.16.

S-type (solid circles) fractionation models are illustrated in an identical fashion as above. Only two representative fractionation curves are shown. Fractionating assemblages comprise plagioclase, biotite and quartz. Note contrasting Ba concentrations for 75-77% SiO₂ S-types and 75-77% SiO₂ I-types. Circles containing crosses denote uncommon I-type suites that display early biotite fractionation signatures i.e. evolved liquids are depleted in Ba.
S-types

- I-types of northern segment (Boundary Range, Kapal, Jengai Batholiths and associated stocks).

- I-types of southern segment (Singapore Complex, Besar and Blumut-Muntahak Batholiths and associated stocks).

---

The high Sr (500 ppm) in the highly silicic samples suggests a significant amount of biotite crystallization (if > 0.5) is required. The model compositions are shown in Fig. 102. Two of the trends (curves 1, 2) are based on the fractionating assemblages estimated by major element models, whereas the third (curve 3) is based on late crystallization of biotite. All three curves display an initial increase of Rb, with decreasing Sr due to plagioclase and hornblende removal. Initiation of biotite crystallization causes the slope to flatten out. Both Sr and Rb decrease in the later stages when biotite and K-feldspar are reached. Only a representative area labelled 'I-type solids' trends is shown as all solid trends produce an identical effect in the
solids from liquids is a concurrently operating mechanism.
Incorporation of significant amounts of interstitial liquid trapped
within the framework of early crystallizing solids to form tonalitic –
mafic granodioritic marginal phases of plutons would leave much less
derivative felsic liquid than the 40% implied by f = 0.6.

Within the constraints imposed by the major element oxide
modelling, the bulk of the Rb – Sr – Ba variations (with the exception
of several high – Sr plutons) can be modelled as resulting by
fractional crystallization processes from original magmas with
estimated LIL element abundances shown below:

Rb = 70–120 ppm

Sr = 250–850 ppm

Ba = 350–700 ppm

To explain the high Ba (>900 ppm) in some highly siliceous granites,
late initiation of biotite crystallization (f ≥ 0.4) is required
whereas hornblende and biotite can crystallize together over a large
range of f values (0.2 to 0.6) for most samples.

Ba – Sr : Three calculated liquid trends for three different initial
compositions are shown in Fig.18A. Two of the trends (curves 1,2) are
based on the fractionating assemblages estimated by major element
models whereas the third (curve 3) is based on late crystallization of
biotite. All three curves display an initial increase of Ba with
decreasing Sr due to plagioclase and hornblende removal. Initiation
of biotite crystallization causes the slope to flatten out. Both Sr
and Ba decrease in the later stages when biotite and K-feldspar are
removed. Only a representative area labelled 'I-type solids' trends
is shown as all solid trends produce an identical effect ie. the
incorporation of crystallized solids will push magma compositions to higher Sr - lower Ba fields. Incorporation of restite will produce similar results.

Ba - Rb: Both Ba and Rb, being incompatible with respect to initial fractionating phases, increase until biotite crystallization begins. The initially steep positive trends flatten out and finally develop negatively sloped 'tails' when biotite and K-feldspar coprecipitate. Incorporation of crystals or restite shifts magma compositions to low Ba - low Rb fields i.e. lower left corner of diagram.

Rb - Sr: All trends show smooth curves of increasing Rb with decreasing Sr. Incorporation of early crystals or restite shift magma compositions to high Sr - low Rb fields.

Four unrelated samples characterized by high Sr (400 - 800 ppm) cannot be explained by the general model discussed above using the LIL element source abundances listed previously. These samples do not possess cumulate characteristics and lie outside the region of possible cumulate compositions for the other samples of the province. They appear to belong to a distinct group of high-Sr granitoid plutons.

To summarize the main conclusions, (1) a general crystal fractionation-unmixing mechanism was presented to explain the major element oxide and Ba - Rb - Sr variations of the East Coast Province I-type plutons, and (2) broad compositional characteristics of the parental liquids in terms of SiO₂, Rb, Ba and Sr were suggested. The contrasting fractionation patterns of the East Coast Province S-type magmas that are dominated by biotite + plagioclase fractionation signatures as compared to early plagioclase + hornblende for I-type
magmas are discussed next.

**B) S-type fractionation processes**

Early biotite fractionation is characteristic of, but not uniquely restricted to S-type granitoids and the trends that will be described are also shown by at least two I-type plutons. The field of these I-type plutons are, however, clearly separated from the S-type field even though the trends are similar. This essentially reflects the different source compositions of the two granitoid types. The methodology adopted for modelling is identical to that described previously for the I-type granitoids. Hornblende can be excluded as a fractionating phase for S-type granitoids. Minerals fractionated are dominated by plagioclase, biotite, quartz and K-feldspar. A typical permitted solution to the major element trends is listed in table 7A-B. The fractionating assemblages at various stages comprise 40 - 55% plagioclase, 10 - 25% biotite, 15 - 30% quartz and up to 25% K-feldspar. Quartz is required as an early crystallizing phase for major element oxide models to be acceptable. K-feldspar is suggested as a fractionating phase after ~10% solidification not because it is necessary for successful major element oxide modelling per se but because both biotite and K-feldspar fractionation is required in order to achieve the extremely low Ba values shown by some high silica granites.

The general Ba, Sr and Rb variations (Fig.18A-C) can be explained by concurrent crystal fractionation - unmixing processes from original melts with estimated LIL element abundances shown below:
Table 7A Fractionating phases utilised for modelling of S-type trends

<table>
<thead>
<tr>
<th>Mineral Phase</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>0.55</td>
<td>0.53</td>
<td>0.50</td>
<td>0.47</td>
<td>0.44</td>
<td>0.42</td>
<td>0.40</td>
</tr>
<tr>
<td>biotite</td>
<td>0.25</td>
<td>0.21</td>
<td>0.18</td>
<td>0.18</td>
<td>0.16</td>
<td>0.14</td>
<td>0.11</td>
</tr>
<tr>
<td>quartz</td>
<td>0.19</td>
<td>0.20</td>
<td>0.23</td>
<td>0.25</td>
<td>0.27</td>
<td>0.27</td>
<td>0.27</td>
</tr>
<tr>
<td>K-feldspar</td>
<td></td>
<td>0.05</td>
<td>0.085</td>
<td>0.115</td>
<td>0.145</td>
<td>0.20</td>
<td>0.25</td>
</tr>
<tr>
<td>Ilmenite</td>
<td>0.01</td>
<td>0.01</td>
<td>0.005</td>
<td>0.005</td>
<td>0.005</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

* denotes 10% by mass of preceding liquid, not original liquid

Table 7B Typical evolving S-type liquid compositions resulting from major element oxide modelling

<table>
<thead>
<tr>
<th>Oxide</th>
<th>0</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>70.0</td>
<td>71.3</td>
<td>72.4</td>
<td>73.5</td>
<td>74.3</td>
<td>74.9</td>
<td>75.3</td>
<td>75.7</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.5</td>
<td>14.0</td>
<td>13.5</td>
<td>13.1</td>
<td>12.8</td>
<td>12.4</td>
<td>12.1</td>
<td>11.8</td>
</tr>
<tr>
<td>FeO</td>
<td>2.5</td>
<td>2.1</td>
<td>1.75</td>
<td>1.4</td>
<td>1.0</td>
<td>0.7</td>
<td>0.4</td>
<td>0.2</td>
</tr>
<tr>
<td>MgO</td>
<td>0.6</td>
<td>0.47</td>
<td>0.37</td>
<td>0.29</td>
<td>0.22</td>
<td>0.17</td>
<td>0.13</td>
<td>0.1</td>
</tr>
<tr>
<td>CaO</td>
<td>2.2</td>
<td>1.85</td>
<td>1.58</td>
<td>1.31</td>
<td>1.13</td>
<td>0.95</td>
<td>0.89</td>
<td>0.83</td>
</tr>
<tr>
<td>K₂O</td>
<td>4.0</td>
<td>4.2</td>
<td>4.4</td>
<td>4.6</td>
<td>4.8</td>
<td>5.0</td>
<td>5.2</td>
<td>5.4</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.5</td>
<td>3.6</td>
<td>3.6</td>
<td>3.6</td>
<td>3.5</td>
<td>3.4</td>
<td>3.2</td>
<td>3.0</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.4</td>
<td>0.32</td>
<td>0.23</td>
<td>0.16</td>
<td>0.09</td>
<td>0.02</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Note: Calculated trends of increasing Si and decreasing K₂O, MgO, Al₂O₃, Na₂O, CaO, TiO₂, and FeO; observed trends of decreasing Si, MgO, Al₂O₃, Na₂O, CaO, TiO₂, and FeO; calculated trends of increasing K₂O, CaO, MgO, and FeO; observed trends of increasing K₂O, CaO, MgO, and FeO. The high silica S-types require extreme low K₂O values (> 80 - 90% solidification) to achieve because of the asymptotic nature of the modelled curves. In view of the feasibility of significant trapped interstitial liquid being held up by early crystallising solids, it is suspected that such extreme fractionation is unrealistic for high-silica
Rb = 200-250 ppm
Sr = 100-200 ppm
Ba = 200 - 400 ppm

These estimates when compared to those predicted for the I-type melts of the East Coast Province in the previous section are low in Ba possibly because biotite coexist with S-type peraluminous melts during crustal melting processes (Clemens and Wall, 1981). The higher Rb/Sr ratios indicated reflect the higher values of this ratio in the metasedimentary source rocks of S-type magmas when compared to the lower values of mantle and lower crustal rocks that are the postulated sources for I-type magmas.

Ba - Sr : Due to early biotite fractionation, modelled Ba - Sr trends are antithetic to those of I-type trends. Both Ba and Sr decrease to very low values in the S-type suites studied here. A 77% SiO₂ S-type may have Ba values < 50 ppm whereas a 77% SiO₂ I-type may have > 1000 ppm Ba yet both are similarly characterized by low Sr values due to plagioclase fractionation.

Ba - Rb : Calculated trends of increasing Rb and decreasing Ba define hyperbolic curves as shown in Fig.18B. One aspect of the Ba-Rb modelling results deserves comment. Although the trends of the S-type plutons can be accounted for by the fractionation-unmixing mechanism, two high silica S-types possess extremely low Ba values (5 ppm and 20 ppm) that require extended conventional fractionation (~80 - 90% solidification) to achieve because of the asymptotic nature of the modelled curves. In view of the inevitability of significant trapped interstitial liquid taken up by early crystallizing solids, it is suspected that such extreme fractionation is unrealistic for granitoid
magnas. Perhaps additional processes like liquid state diffusion as advocated by Hildreth (1981) or reaction with a fluid phase may have occurred. It is stressed that such superimposed processes need not be invoked for the bulk of the other samples. Other samples from the same plutons as these low Ba samples show higher Ba values (50 - 90 ppm) and are adequately explained by fractionation processes. If liquid state diffusion processes operate and their effects are not ephemeral, they are only localized phenomena in the felsic plutons described.

Rb - Sr : The directions of the modelled Rb - Sr trends are similar to those of I-type granitoids i.e. curved trends of increasing Rb with decreasing Sr as fractionation progresses.

Aspects of process identification

Rb, Sr and Ba variations in the S- and I-type granitoid associations of the East Coast Province can be explained in terms of concurrent fractionation - unmixing processes. These LIL element trends best reflect the large scale magma processes that occur. For the sake of clarity, more complex scenarios were ignored as the dominance of the effects of crystal-liquid fractionation equilibria were clearly beyond doubt. In this section, it is appropriate to make certain qualifying remarks with regard to other possible mechanisms.

(a) combined assimilation-fractionation : DePaolo (1981a) formulated an equation to describe simultaneous assimilation - fractional crystallization processes. The equation takes the form:

\[
\frac{C^L}{C^O} = F - z + \left( \frac{r}{r-1} \right) \frac{C_a}{zC^O} (1-F - z)
\]

where \( F = \frac{\text{mass of liquid } ^L}{\text{mass of original liquid}} \); \( z = \frac{r + D - 1}{r-1} \); \( r = \frac{\text{rate of assimilation}}{\text{rate of fractionation}} \);

\( C_a = \text{concentration of element in wall rock, } C^L = \text{concentration of element in a derivative liquid, } C^O = \text{concentration of element} \)
in original liquid; \( D \) = bulk solid/liquid partition coefficient.

Substituting appropriate values used in previous modelling calculations and \( r = 0.05 \) to \( 0.1 \), \( Ca = 350 \) ppm for Ba and 400 ppm for Sr (typical 'crustal' values), it was found that the results obtained are essentially identical to those of the incremental equilibrium fractionation mode for Ba and Sr. Bearing in mind that the Ba versus Sr trends for S-types probably provide the ideal situation to test any differences between closed system fractionation versus combined assimilation-fractionation, it may be concluded that under normal circumstances, simultaneous assimilation - fractionation processes cannot be distinguished from closed system fractionation using Rb, Ba and Sr. Assimilation - fractionation processes, theoretically, result in the achievement of a steady state concentration (\( \sim 30 \) ppm for Ba and Sr for the parameters used). This buffering is only achieved after \( > 90\% \) fractional crystallization. The potential usefulness of this feature is severely limited because most granitoid magmas do not achieve such extended fractionation.

(b) Magma mixing: The clear dichotomy of the two contrasting granitoid associations argues against any general mixing between I- and S-type magmas in the East Coast Province. Intra - S-type or intra - I-type mixing in high level magma chambers is possible but there is a lack of anomalous mineral textures like sodic plagioclase cores rimmed by calcic mantles, K-feldspar grains with plagioclase mantles and highly magnesian amphiboles rimming less magnesian amphiboles. Mixing of magma batches at depth or near source regions followed by ascent and crystal fractionation in high level magma chambers would not detract from any of the conclusions reached based on the LIL element modelling as such 'hybrids' can be viewed as a single 'magma' when fractionation processes dominate at shallow crustal levels.
(c) Fractional melting: Presnall and Bateman (1973) have developed cogent arguments against the process of fractional fusion (i.e. repeated extraction of liquids in small amounts from a single source) to explain calcalkaline suites in granitoid batholiths. Fractional fusion yields liquids that lie on the quartz-saturation surface in the Ab-An-Or-Qtz tetrahedron. As such, the process cannot yield calcalkaline compositions that trend at a high angle to the quartz saturation surface. In addition, fractional fusion cannot produce the extremely low abundances of Ba and Sr in the highly felsic S-type compositions from normal source compositions.

(d) Liquid state differentiation: Hildreth (1981) discussed the mechanisms by which compositional variations in liquids can arise and explained the effects of Soret diffusion, melt structure control of trace element partitioning, thermogravitational fractionation and rare metal complexing. He attributed the chemical characteristics of the rhyolitic units of the Bishop Tuff to the operation of such processes. The characteristics, in particular, the extreme depletion of Sr, Ba, Eu; the LREE depletion and enrichments in Th, U, Rb and Y are so exactly duplicated in the felsic S-type granites (75 - 77% SiO₂) studied here that a discussion is warranted. Felsic I-types do not show these features and their Ba - Sr relationships are antithetic to felsic S-type granitoids. Ba, Sr and Rb trends of the S-type suites are best explained by liquid-crystal fractionation equilibria. Nb and LREE contents of the S-types do not correlate with increasing Th, U as would be expected for transport of lanthanides and actinides as volatile complexes. Low values of the LREE and high U/Th ratios are more consistent with monazite fraction than liquid state differentiation as monazite removes LREE and Th in preference to HREE and U respectively. Thus, characteristics attributed to liquid
differentiation processes can also be consistently explained by liquid-crystal equilibria for the felsic S-type granites. In fact, quartz+K-feldspar + sodic plagioclase can be fractionated from a granitic liquid to yield a liquid that is still granitic in composition but which is severely depleted in Sr, Ba and Eu. It is concluded that liquid state differentiation effects need only be minor effects superimposed on impressive fractionation signatures developed during differentiation from granodiorites to granites for the S-type association. The features displayed by the felsic I-type granites do not conform to liquid state differentiation signatures documented by Hildreth (1981).

Other trace element characteristics

Fig. 19A shows the La vs. Th plot for the East Coast Province granitoids. S-type granitoids, in general, have lower La/Th ratios (0.25 - 1.5) than I-types (1.0 - 6.0). Low La/Th ratios of S-types are probably an inherited feature from the sedimentary source rocks of S-type magmas. Bhatia and Taylor (1980) showed that La/Th ratios of volcanogenic sediments are higher than those of recycled clastic sedimentary rocks in the Tasman Geosyncline of southeastern Australia. A possible reason cited by them is the preferential retention of Th relative to La in clays.

Other notable chemical features include the high Y abundances of S-types and their trend of increasing Y (which behaves similarly to the HREE) with decreasing La (Fig. 19B). Fig. 19C shows the tendency of felsic S-types towards low Th/U ratios (<3.0) whereas mafic S-types and both mafic and felsic I-types display typical Th/U ratios of 4 to 6. As pointed out previously such trends may be attributed to monazite fractionation.
Fig. 19. A: plot of La vs. Th for the East Coast Province granitoids; B: Plot of Y vs. La; C: Plot of U vs. Th. Symbols as for Fig. 18.
With regard to the REE, both S- and I-types are typified by LREE-enriched chondrite normalized patterns (Figs. 20A-B). The limited data show that felsic S-types are characterized by high HREE and lower LREE relative to more mafic S-types. These combined effects lead to the development of profiles approaching flat REE patterns in felsic S-types. All S-types analysed possess negative Eu anomalies. These anomalies are emphasized to an extreme degree in felsic S-types due to plagioclase and K-feldspar fractionation.

I-types have consistently higher LREE and lower HREE than S-types. Although major element oxide and LIL element models show that plagioclase is just as important a fractionating phase compared to S-types, negative Eu anomalies are relatively suppressed, even in the most felsic I-type granites. Two of the most mafic I-type plutons lack Eu anomalies. There is no systematic correlation of SiO content with increasing LREE or HREE abundances.

Eu can exist in both $2^+$ and $3^+$ states. Eu in the $2^+$ state can be expected to be more important in reduced magmas. Ilmenite composition data and sphene + magnetite assemblages in I-type plutons suggest that I-type magmas are more oxidized and S-type magmas more reduced. Because of the similar size and charge of $\text{Sr}^{2+}$ (1.18A) and $\text{Eu}^{2+}$ (1.25A) compared to $\text{Eu}^{3+}$ (0.98A), $\text{Eu}^{2+}$ and $\text{Sr}^{2+}$ will exhibit parallel behaviour in reduced S-type magmas. In this sense, the extreme depletion of Sr would correlate well with the extreme negative Eu anomalies shown by the felsic S-type plutons in the East Coast Province. Fractionation of K-feldspar would further facilitate depletion of both Eu (Leeman and Phelps, 1981) and Sr abundances in S-type magmas compared to I-type magmas as K-feldspar is not an important fractionating phase in I-type magmas. The extreme negative
Table 8  REE abundances of representative I- and S-type granitoids

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Fig. 20. A-B. Chondrite - normalized REE plots of selected S- and I-type granitoids. Sample numbers are followed by SiO₂ contents of samples in parentheses for each figure.
Eu anomalies of the S-type leucogranites may thus be the result of both the reduced nature of S-type magmas and the importance of plagioclase + K-feldspar fractionation.

ROCKS OF THE MAFIC BATHOLITH ASSOCIATION

Mafic rocks associated with the Permian to Triassic batholiths of the East Coast Province constitute less than five percent of the total exposed area of the batholiths. They are found as early complexes marginal to granitoid plutons, as synplutonic dykes and as late intrusive dykes. Three important mafic intrusive complexes have been mapped. Each intrusive complex is characterized by its distinct lithologic association. The three complexes (Fig.1) have been described by Hutchison (1964), Kumar (1980, 1981) and others. Generalized lithological associations include norite-orthopyroxenite (Singapore mafic stock), olivine eucrite (Linden stock), and gabbro-norite-diorite (Kemuning stock). Chemical and isotopic data for selected samples of these mafic complexes are discussed in this section.

Singapore mafic stock (samples 89, 99): The dominant rock type is a hornblende norite. Sample 89 is typical and contains phenocrysts of hypersthene (En$_{53-55}$) and bytownite (An$_{70-80}$) in a medium-granied matrix of labradorite (An$_{67}$), hypersthene (En$_{47}$), augite (Mg$_{35}$Fe$_{19}$Ca$_{46}$), interstitial magnesio-hornblende (Mg$_{42}$Fe$_{31}$Ca$_{27}$), biotite (Mg$_{47}$) and quartz. Ilmenite (Ilm$_{96}$) is the sole opaque oxide. Opx + plag + cpx + amph + biotite is the inferred crystallization sequence. Sample 99 is an orthopyroxenite consisting of cumulus hypersthene - bronzite (En$_{68-71}$) set poikilitically in a mosaic of colourless magnesio-cummingtonite (Mg$_{66}$Fe$_{30}$Ca$_{4}$), brownish
Table 9  Microprobe analyses of mafic silicates in mafic samples

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1  total Fe as FeO ; 2 : Mg²⁺ - atomic (Mg/Mg+Fe)

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magnesio-hornblende \((\text{Mg}_{56}\text{Fe}_{17}\text{Ca}_{27})\), bytownite \((\text{An}_{87})\), phlogopite \((\text{Mg}_{70})\) and quartz. Much of the hornblende is replacing cummingtonite. Interstitial ilmenite \((\text{Ilm}_{99})\) is the sole opaque oxide. The inferred crystallization sequence is opx \(\rightarrow\) Ca-poor amphibole \(\rightarrow\) plag \(\rightarrow\) Ca-rich amphibole \(\rightarrow\) phlogopite.

**Synplutonic dyke** (sample 91): Mafic dykes intrusive into a granodiorite pluton are common on Pulau Ubin, an island lying off northeast Singapore. Sample 91 is a sample from a sinuous synplutonic dyke showing chilled margins and boudinaged features. In thin-section, it displays a hornfelsic texture. Minerals present are andesine \((\text{An}_{48})\), granular aggregates of optically continuous diopsidic augite \((\text{Mg}_{37}\text{Fe}_{19}\text{Ca}_{44})\) and hypersthene \((\text{En}_{53})\), biotite \((\text{Mg}_{54})\) and magnesio-hornblende \((\text{Mg}_{43}\text{Fe}_{29}\text{Ca}_{28})\). Ilmenite \((\text{Ilm}_{99})\) and iron sulphides are the opaque phases. Accessory apatite is abundant.

**Linden stock** (samples 96, 97): Eucrites with variable proportions of orthopyroxene and clinopyroxene and often olivine-bearing make up the Linden stock. Sample 96 consists of olivine \((\text{Fo}_{73-76})\), laths of bytownite-anorthite \((\text{An}_{88-92})\), intercumulus orthopyroxene \((\text{En}_{75-77})\) and intercumulus clinopyroxene \((\text{Mg}_{46}\text{Fe}_{10}\text{Ca}_{44})\). Ilmenite \((\text{Ilm}_{99})\) occurs as scattered anhedral grains. Green spinel is present but is only associated with patches of secondary amphibole. Sample 97 is an olivine-free eucrite with bytownite \((\text{An}_{80})\) and altered pyroxenes. The Linden rocks are quite typical plag+opx+cpx+olivine cumulates showing a crystallization sequence similar to oceanic tholeiites i.e. olivine + plag + cpx + opx.
Kemuning stock: Kumar (1980) described a suite of high Ti-ferrogabbros containing abundant hornblende, biotite and apatite. The inferred crystallization sequence is oliv+plag→cpx→opx→hydrous phases. Olivine and orthopyroxene compositions range from Fo$_{32}$–Fo$_{14}$ and En$_{53}$–En$_{37}$ respectively. Augite–ferroaugite is the coexisting clinopyroxene. Plagioclase ranges from An$_{72}$ to An$_{50}$.

Notable mineralogical features of the subalkaline mafic rocks documented here are the ubiquity of orthopyroxene and, with the exception of the Linden eucrites, the abundance of hornblende and biotite. Another characteristic is the very calcic plagioclase coexisting with olivine of a particular forsterite value and orthopyroxene of a particular enstatite value (Fig. 21). This characteristic is a consistent feature of island and continent margin arc mafic rocks. It has been attributed to the increased activity of H$_2$O in such rocks which depresses the stability of plagioclase to lower temperature more than pyroxene or olivine (e.g. Arculus and Wills, 1980). High water contents is also implied by the abundance of hornblende and biotite observed even in primitive rock compositions.

Coexisting pyroxene compositions are displayed on the pyroxene quadrilateral in Fig. 22. Temperatures estimated using the equation of Wells' (1977) and experimental compositional data of Ross and Huebner (1975) agree well with each other and lie in the range 930–1100°C. The reason for the ubiquity of orthopyroxene is not certain although one possibility is high a$_{SiO_2}$ (Powell, 1978, p. 209) in these mafic magmas. High a$_{SiO_2}$ shifts the direction of the equilibrium

\[
\text{Mg}_2\text{SiO}_4 + \text{SiO}_2 \rightarrow \text{Mg}_2\text{Si}_2\text{O}_6 \\
\text{olivine melt} \quad \text{orthopyroxene}
\]

to the right.
Fig. 21. Plot of % An in plagioclase vs. % En in coexisting orthopyroxene for rocks of the mafic association. Incorporates data from Kumar (1980). Arc suites selected for comparison are Lesser Antilles (LA)(Arculus and Wills, 1980, and Arculus, pers. comm.), Uasilau - Yau Yau (UYY)(Whalen, 1980) and Bear Mt. Complex (BM)(Snoke et al., 1981). Data for DSDP site 334 and stratiform complexes are from Hodges and Papike (1976), Papuan Ultramafic Belt (PUB) are after Jaques and Chappell (1980) and Bay of Islands ophiolite are after Church and Riccio (1977).
Fig. 22. Plot of coexisting pyroxenes on the pyroxene quadrilateral. Isotherms in °C are those of Ross and Huebner (1975). Unnumbered samples are data from Kumar (1980) for the Kemuning Complex.
The mafic rocks analysed range from 45% to 55% SiO$_2$ (Table 10). Minor dioritic phases within the mafic complexes may reach 60% SiO$_2$ compositions (Kumar, 1980, 1981). Chemical arguments like low Fo contents of olivine, low En contents of orthopyroxene suggest that even the most primitive of these rocks have crystallized from magmas that have undergone some degree of prior fractionation. Many workers have shown that partial melting of mantle lherzolite should yield tholeiitic and picritic liquids with Mg-numbers ~70-74, high Ni (250 - 300 ppm), high Cr (500 - 600 ppm), and crystallize early olivine ~ Fo$_{90}$ and orthopyroxene ~ En$_{88}$ (e.g. Duncan and Green, 1980; Perfit et al., 1980b; Gill, 1981).

Major element oxide characteristics of the mafic complexes are shown on an AFM diagram (Fig.23) and a TiO$_2$ vs. MgO/(MgO+FeO) plot (Fig.24). Separate fields for each complex can be discerned. There is a general lack of extended fractionation to felsic compositions near the Na$_2$O $+$ K$_2$O corner on the AFM diagram. Both plots also highlight the high Fe - high Ti characteristics of the Kemuning suite.

Singapore stock: Mineralogical features and chemical arguments like persistently high SiO$_2$ contents (~53%), decreasing FeO, MgO with increasing CaO, Na$_2$O, Al$_2$O$_3$ and similar ratios of immobile incompatible elements (Zr/Nb, La/Th, Zr/Y) suggest that samples 89 and 99 are related by orthopyroxene fractionation. Of particular interest is the nature of the parental magma for these rocks implied by the nature of the orthopyroxenite (sample 99). The unusual mineral assemblage of this rock (cumulus bronzite $+$ An$_{90}$ plagioclase $+$ cummingtonite $+$ Mg-hornblende $+$ phlogopite $+$ quartz) results in a distinctive rock chemistry: high MgO (18.7%), very low CaO (3.9%),
Table 10  Major and trace element analyses of representative samples of mafic rocks.

<table>
<thead>
<tr>
<th>Sample</th>
<th>SIO₂</th>
<th>TIO₂</th>
<th>AL₂O₃</th>
<th>Fe₂O₃</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>SiO₂⁺</th>
<th>H₂O⁻</th>
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<tbody>
<tr>
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<td>0.94</td>
<td>17.28</td>
<td>1.33</td>
<td>7.65</td>
<td>0.14</td>
<td>5.12</td>
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<td>0.16</td>
<td>0.01</td>
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<td>16.50</td>
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<td>0.15</td>
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<td>10.09</td>
<td>2.92</td>
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<td>0.50</td>
<td>0.04</td>
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<td>0.11</td>
<td>5.83</td>
<td>13.81</td>
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<td>0.41</td>
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<td>18.95</td>
<td>99.97</td>
<td>99.95</td>
<td>99.37</td>
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Trace elements:

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<th>MAL96</th>
<th>MAL97</th>
<th>MAL99</th>
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<td>5</td>
<td>7</td>
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<td>Th</td>
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<td>&lt;0.2</td>
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<td>1.5</td>
<td>1.5</td>
</tr>
<tr>
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</tr>
<tr>
<td>Zn</td>
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<td>Ga</td>
<td>18.4</td>
<td>19.2</td>
<td>12.2</td>
<td>15.4</td>
<td>6.8</td>
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Fig. 23. AFM plot of mafic rocks of the East Coast Province to illustrate their tholeiitic affinities (Kumar, 1980, 1981).

Fig. 24. Plot of % TiO₂ vs. MgO/(MgO + FeO^t). Field for island arc basalts is after Perfit et al. (1980 b). Data from Kumar (1980, 1981).
low Al₂O₃ (4.5%) and low TiO₂ (0.5%). It also has high contents of elements compatible in basaltic magmas like Cr (670 ppm), Ni (150 ppm), V (250 ppm) and Sc (45 ppm). This accords well with its primitive cumulate nature but more significantly, the mineralogical association and chemical characteristics have much in common with those of high-Mg andesitic island arc lavas called boninites i.e. orthopyroxene fractionation, very calcic late plagioclase, high SiO (~55% SiO₂), high MgO (10-16%), high Cr (>500 ppm), low CaO (4-8%), low Al₂O₃ (< 10%)(Cameron et al., 1979, 1980; Sun and Nesbitt, 1978; Jenner, 1981). The uniqueness of the Singapore body is its occurrence as a plutonic body in a continent margin orogenic batholith setting.

**Synplutonic dyke** (sample 91): The evolved mineral chemistry of sample 91 is corroborated by the major and trace element whole rock chemistry (Table 10) that suggests that extensive prior fractionation had occurred. Its relatively high total alkali content (Na₂O + K₂O = 4.0%), together with high P₂O₅, Ba, Rb, Sr, Zr, Nb and LREE best resemble medium and high-K arc tholeiitic basalts (Gill, 1981).

**Linden stock**: Both samples 96 and 97 show typical cumulate gabbro chemistry. High Sr, Al₂O₃, CaO, positive Eu anomalies (discussed under REE section) and extremely low incompatible element abundance do not represent realistic liquid compositions. Ba, Rb and Sr abundances are however significantly higher than similar pyroxene-plagioclase + olivine cumulates from ophiolites (e.g. Jaques and Chappell, 1980). Expected gabbro-diorite with more evolved compositional features are not known to be associated with these cumulate rocks and this aspect distinguishes the Linden stock from other mafic complexes of the East Coast Province.
Kemuning stock: This high-Ti, ferrogabbro-diorite complex displays many chemical similarities to the synplutonic dyke discussed earlier. For a silica range of 45-60%, total alkalis (Na$_2$O + K$_2$O) range from 3-5% (Kumar, 1980). In addition to high alkalis, the rocks also show LILE-enrichment characteristics of medium to high-K arc rocks. These include high P$_2$O$_5$, Ba, Rb, Sr, Zr, Th and LREE abundances.

Chondrite-normalized REE patterns of the mafic rocks are shown in Fig.25. All samples show LREE-enriched patterns. Both samples from the Singapore stock display intriguing concave upward patterns. This is especially evident in the primitive orthopyroxenite sample which, in addition, possesses a negative Eu anomaly. Such concave patterns are also shown by high-Mg andesitic rocks and ophiolitic basalts. The mechanism discussed by Green (1976) and Sun and Nesbitt (1978) to explain such patterns is the addition of LREE-enriched components to an otherwise depleted source during melting. I have accepted this mechanism to be a viable one and have attributed the Eu anomaly as an artifact of this mixing mechanism rather than a source characteristic. The less evident concave pattern of the more evolved norite (sample 89) is the result of enrichment of the incompatible light and middle REE during fractionation.

The Linden olivine-eucrite cumulate possesses low abundances of the REE (La ~ 3x chondritic). The prominent positive Eu anomaly and low abundances of the REE are consistent with the primitive cumulate nature of sample. Unlike gabbroic cumulates from ophiolites (Jaques and Chappell, 1980; Pallister and Knight, 1981) which have characteristic LREE-depleted patterns similar to MORB, the Linden eucrite is typified by a LREE-enriched pattern. In this respect, it is similar to other gabbros found in orogenic terrains (e.g. Bear
Table 11  REE, Hf, Ta and Ba abundances of mafic samples

<table>
<thead>
<tr>
<th></th>
<th>89</th>
<th>91</th>
<th>96</th>
<th>99</th>
<th>145</th>
<th>146</th>
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<tbody>
<tr>
<td>La</td>
<td>10.57</td>
<td>54.51</td>
<td>1.06</td>
<td>3.64</td>
<td>56.20</td>
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</tr>
<tr>
<td>Ce</td>
<td>24.27</td>
<td>111.40</td>
<td>2.73</td>
<td>7.91</td>
<td>102.66</td>
<td>42.20</td>
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<tr>
<td>Nd</td>
<td>13.56</td>
<td>49.30</td>
<td>1.87</td>
<td>3.49</td>
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<td>23.90</td>
</tr>
<tr>
<td>Sm</td>
<td>3.23</td>
<td>8.77</td>
<td>0.49</td>
<td>0.90</td>
<td>10.45</td>
<td>4.50</td>
</tr>
<tr>
<td>Eu</td>
<td>1.08</td>
<td>2.20</td>
<td>0.38</td>
<td>0.21</td>
<td>2.88</td>
<td>1.70</td>
</tr>
<tr>
<td>Gd</td>
<td>3.73</td>
<td>7.14</td>
<td>0.69</td>
<td>1.11</td>
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<td>-</td>
</tr>
<tr>
<td>Tb</td>
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<td>Lu</td>
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<tr>
<td>Hf</td>
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<td>5.1</td>
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<td>0.55</td>
<td>3.1</td>
<td>1.30</td>
</tr>
<tr>
<td>Ta</td>
<td>0.44</td>
<td>1.05</td>
<td>0.05</td>
<td>0.10</td>
<td>1.6</td>
<td>0.75</td>
</tr>
</tbody>
</table>
Fig. 25. Chondrite-normalized REE plot of representative samples from the East Coast Province mafic complexes. Samples 145, 146 are samples from the Kemuning Complex kindly supplied by S.C. Kumar. These correspond to samples M13C and KE 31D of Kumar (1980). 145 is a hornblende gabbro-norite and 146 is a quartz-biotite-hornblende gabbro-norite.
Mountain Complex (Snoke et al., 1981)).

Mafic rocks suggested to be members of a distinctive medium-K tholeiitic association (Kemuning samples 145, 146; synplutonic dyke 91) display steep LREE-enriched patterns and very high abundances of the LREE (La ∼ 80 - 200X chondritic). No Eu anomalies are discernible.

A useful way to describe trace element variations of mafic rocks is to compare them with respect to a reference composition. Hypothetical primordial mantle compositions are useful in this context because they yield information pertaining to melting and fractionation events involved in generating present-day mantle-derived magmas. Wood et al. (1979) proposed a diagram that utilized a model primordial mantle normalization and suggested values for incompatible trace element abundances of primordial mantle calculated from estimates of bulk earth composition.

Elemental abundances of mafic rocks are displayed on a Wood et al. (1979) diagram by plotting values normalized to primordial mantle estimates on a logarithmic scale along a vertical axis and the various incompatible elements plotted along a horizontal axis. Elements are arranged in order of increasing D (solid-liquid partition coefficient) values for realistic mantle mineralogies using either experimentally determined mineral-liquid Kd's or estimated from relative enrichment factors in a single basalt fractionation series. Fig. 26 shows the patterns of the East Coast Province mafic samples on such a normalized plot. A typical MORB pattern is also shown as a reference. Patterns obtained for the studied samples are characteristic of arc basalts (e.g. Perfit et al., 1980b). The more important of these are:
Fig. 26. Plot of incompatible element abundances of representative mafic rocks of the East Coast Province normalized to primordial mantle abundances estimated by Wood et al. (1979). See text for discussion.
1. the overabundances of the LIL elements (K, Rb, Cs, Th, Ba, U) relative to the HFS (high field strength) elements (Hf, Zr, Ti, Y, P) and HREE,

2. the unsystematic manner of the LIL element abundances, and

3. common negative Nb and Hf 'spikes'.

Spikes for Sr and Zr displayed by the Linden eucrites are a result of their cumulate nature and are not source characteristics. High abundances of LIL elements relative to REE and HFS elements described for arc basalts can also be illustrated on conventional variation diagrams. For example, a La vs. Th (Fig. 27) plot shows that arc basalts are distinguished by low La/Th ratios.

Discussion

The mafic complexes discussed display broad geochemical affinities with the orogenic tholeiitic magma series (Gill, 1981). Each complex, however, shows its own distinctive mineralogical and chemical characteristics. The complexes cannot be related genetically by fractionation to any of the voluminous granitoid intrusions of the East Coast Province. Timing of mafic magmatism during batholith constitution range from pre-granitoid and syn-granitoid to post-granitoid emplacement. Such intimate spatial and temporal relationships between genetically unrelated mafic and felsic magmas is compatible with the intrusive mechanism proposed by Pitcher (1978) who suggested that deep zones of weakness are utilised by various magmas to facilitate their ascent. Despite their minor volumes, the significance of the presence of the mafic complexes is that they may be representative of the 'mantle heat trigger' that initiates
Fig. 27. Plot of La vs. Th for representative mafic samples. Figure is from Gill, 1981, p. 139; as are fields for high-K, medium-K and low-K orogenic andesites.
widespread melting at lower crustal depths to produce granitoid magmas (Presnall and Bateman, 1973; Hildreth, 1981).

Trace element relationships like those displayed in Fig.26 are rather diagnostic of arc magmas. In particular, the high LIL/HFS element ratios indicate an unsystematic enrichment of highly mobile LIL elements. Such an explanation favours an argument for the external influx of volatiles enriched in LIL elements that suppress mantle solidii sufficiently for partial melting to occur. Saunders et al. (1980), for example, suggested that a heterogenous mantle wedge containing variable amounts of a LIL element - enriched vein component derived from the subducting slab removes the problem of accounting for the high LIL element abundances in models of fusion of lherzolitic mantle at subduction settings. They further suggested that stabilisation of accessory phases like sphene, ilmenite, rutile and apatite under high P\textsubscript{H2O} conditions would further deplete the HFS elements of any coexisting melt. Negative 'spikes' observed for Nb, Ta and Hf (Fig.26) seems compatible with rutile, ilmenite and zircon retention.

The ubiquity of orthopyroxene, absence of magnetite, high FeO/MgO ratios and iron-enrichment (Kemuning suite, synplutonic dyke 91) suggest that parental magmas were fundamentally tholeiitic - probably high-Al\textsubscript{2}O\textsubscript{3} olivine tholeiite for the Linden eucrites, hydrous high-Mg quartz tholeiite for the Singapore suite and hydrous medium-K tholeiitic basalt for the Kemuning suite and synplutonic dyke sample 91. To highlight the Fe- and Ti- enriched characteristics and high alkalis, LREE, P\textsubscript{2}O\textsubscript{5}, LIL element abundances of the Kemuning suite and the synplutonic dyke, these rocks have been described as showing characteristics of a medium-K tholeiite fractionation series.
However, the arbitrary division of rock suites into tholeiitic, calcalkaline and alkalic series based on alkali contents, iron-enrichment and mineralogy is fraught with difficulties (e.g. Gill, 1981) and may even be not meaningful (Johnson et al., 1978). Kumar (1980) feels that the Kemuning suite display chemical and mineralogical features that do not uniquely place it in any single category.

**ISOTOPIC RELATIONSHIPS**

The Nd and Sr isotopic data referred to in this section are described in more detail in chapter six. In this chapter, only aspects of the isotopic results relevant to the petrogenesis of the batholiths of the East Coast Province will be discussed.

### Sr initial ratios

Plutons of the I-type association have Sr initial ratios ranging from 0.705 to 0.710 but are typically around 0.707 - 0.708. Rocks of the mafic association possess ratios ranging from 0.706 to 0.709 and as such, fall entirely within the range of the I-type granitoids. The initial ratios of the mafic rocks are high compared to most primitive arc basalts (e.g. Perfit et al., 1980b) but are similar to values exhibited by basalts and andesites erupted in a number of arc settings (e.g. Indonesian arcs, Whitford et al, 1981; central Andes, James, 1982). Peraluminous S-type suites of the East Coast Province have Sr initial ratios of 0.709 to 0.715. A plot of Sr initial ratio vs. SiO₂ content is shown in Fig.28.
Table 12 - Initial Nd and Sr ratios for plutons of the East Coast Province

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<th>εSr(I)</th>
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Fig. 28. Plot of initial $^{87}\text{Sr}/^{86}\text{Sr}$ vs. % SiO$_2$ for selected samples from the East Coast Province.
**Nd initial ratios**

Nd initial ratios referred to here are expressed using the $\varepsilon$-notation (Initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are displayed in terms of deviations in $10^4$ from the bulk earth value). The most prominent aspect of the Nd isotopic characteristics of the I-type plutons is the large range of $\varepsilon_{\text{Nd}}$ values displayed: from $-0.7$ to $-6$. Rocks of the mafic and S-type association show more restricted variations, $-2.5$ to $-4.5$ and $-3.5$ to $-5.6$ respectively. The absence of any correlation with $\text{SiO}_2$ content of the rocks is shown in Fig. 29.

**Nd - Sr interrelationships**

On an $\varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}}$ diagram (Fig. 30), all the East Coast Province data fall in the lower right hand quadrant, below the bulk earth point. Points representing the mafic rocks do not fall on the 'mantle array' (Depaolo and Wasserburg, 1979) defined by MORB, oceanic island basalts and the bulk earth value. The enriched signatures (i.e. $-\varepsilon_{\text{Nd}} ^{+8}$ + $\varepsilon_{\text{Sr}}$) of the mafic rocks contrast with the depleted signatures of primitive island arc basalts ($\varepsilon_{\text{Nd}} ^{+8}$), MORB ($\varepsilon_{\text{Nd}} ^{+10}$) and gabbros and diorites of the Peninsular Ranges Batholith of California ($\varepsilon_{\text{Nd}} ^{+3}$ to $^{+8}$) (DePaolo, 1981b). This suggests that the LREE - enriched patterns (previous section) of the mafic rocks of the East Coast Province are not solely the result of LREE additions during melting events but also reflect extended time-integrated LREE enrichment of the mantle sources. In this sense, depleted sources like the subducted slab with MORB characteristics ($\varepsilon_{\text{Nd}} ^{+10}$) or mantle regions with characteristics of primitive island arc basalts ($\varepsilon_{\text{Nd}} ^{+8}$) cannot represent realistic sources for the mafic magmas of the East Coast Province. The Nd and Sr data, together with chemical arguments based on the overabundances of LIL
Fig. 29. Plot of initial $\varepsilon_{\text{Nd}} (T)$ vs. % SiO$_2$ for selected samples from the East Coast Province.

Fig. 30. $\varepsilon_{\text{Nd}} (T)$ vs. $\varepsilon_{\text{Sr}} (T)$ plot for granitoids and mafic rocks of the East Coast Province.
elements relative to HFS elements, suggest an enriched mantle source—possibly the mantle wedge pervasively veined with variable amounts of LIL element—enriched materials. (Saunders et al., 1980).

The field of the S-type plutons is distinguished by the large range of $\varepsilon_{\text{Sr}}$ values and a more restricted range of $\varepsilon_{\text{Nd}}$ values. These features yield the rather characteristic subhorizontal field displayed by other S-type batholith terrains (e.g. McCulloch and Chappell, 1982; chapter six). The subhorizontal array can be explained in terms of changes occurring during sedimentary processes. Enrichment of Rb relative to Sr in clays during weathering cycles result in high Rb/Sr ratios and consequently, enhanced evolution of $^{87}\text{Sr}/^{86}\text{Sr}$ with time. In contrast to enhanced and variable $\varepsilon_{\text{Sr}}$ values, $\varepsilon_{\text{Nd}}$ variations of originally differing components tend to be homogenized by sedimentary processes. In this sense, $\varepsilon_{\text{Nd}}$ values of S-type granitoids would be expected to yield a fairly good image of the small $\varepsilon_{\text{Nd}}$ variations of the local metasedimentary basement.

Compared to the S-type field, the field of the I-type plutons on the $\varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}}$ diagram is characterized by a larger range in $\varepsilon_{\text{Nd}}$ values and a range of lower $\varepsilon_{\text{Sr}}$ values. By itself, the isotopic data cannot distinguish between several plausible modes of origin of the I-type magmas—a two reservoir mixing model involving mantle-derived magmas and lower crustal high-grade rocks; a two reservoir mixing model involving mantle-derived magmas and metasedimentary crustal sources represented by the S-type magmas; a three reservoir mixing model involving mantle, lower crustal high-grade rocks and lower grade metapelitic sources and lastly, a single reservoir model involving derivation from old lower crustal and juvenile underplated lower crustal metamorphic lithologies.
Geological processes will evidently blur such end-member scenarios but these remain useful conceptual models for discussion.

The bimodal separation of mafic and granitoid associations and the dichotomy of S- and I-type granitoid associations in the East Coast Province based on chemical and mineralogical arguments suggest that the origin of the I-types is most appropriately explained by the fourth scenario. It is evident that the tripartite division of magma associations in the East Coast Province is possible due to the lack of hybridization, hence the scarcity of intermediate compositions, between basaltic mantle-derived magmas and I-type granodioritic lower crustal-derived magmas. This is quite unlike features of, for example, the Coastal Batholith of Peru (McCourt, 1981) or the Peninsular Ranges Batholith of California (Silver et al., 1979) where tonalites and diorites are dominant rock types and there are no distinct compositional breaks between mafic and felsic rock suites. The implication that extensive hybridization of mantle-and crustal-derived magmas and further fractionation of such hybrids have occurred in the deep source regions of such batholiths is obvious. In this sense, the sources of I-type magmas may be multiple with highly variable degrees of hybridization varying from one plutonic terrain to another. Useful constraints are provided in the East Coast Province by the lack of correlation in $\varepsilon_{\text{Nd}}$ versus $\text{SiO}_2$ and initial $^{87}\text{Sr}/^{86}\text{Sr}$ versus $\text{SiO}_2$ plots (Figs. 28-29). These patterns indicate that it is not likely that the I-types could have been produced solely by mafic magmas undergoing fractionation-assimilation processes that resulted in mixing of mantle-derived magmas and siliceous continental crustal contaminants.
One complexity involves the melting of immature volcanogenic sediments to yield granitoids with I-type characteristics. This possibility has not been considered for the East Coast Province plutons in view of strong isotopic and regional geologic evidences for presence of an established underlying crust of Precambrian age and evidence that such sedimentary-derived 'I-types' are restricted to fold belts dominated by trench-complex sedimentary piles like the New England Fold Belt of eastern Australia (Shaw and Flood, 1981). In this fold belt, S-types are isotopically as immature as associated I-types (McCulloch et al., 1982).

SYNTHESIS
(1) Geotectonic implications

The broad findings of this chapter - the linear batholiths of I-type granitoids, the similarities to continent margin orogenic granitoid batholiths and the 'orogenic' chemical characteristics of the mafic plutonic association are entirely consistent with a subduction-associated origin for the magmas of the East Coast Province. The highly silicic, high K₂O, low CaO chemistries of the I-type suites, the presence of significant volumes of S-type suites, the crustal $\varepsilon_{Nd}$ and $\varepsilon_{Sr}$ signatures of many plutons and zircon inheritance phenomena all suggest that the plutonism was generated at a continental margin plutonic arc setting. Based on Nd model ages and inherited zircon ages (chapter six), it is suggested that a Precambrian continental block was involved in the plutonism.
(2) **Batholith constitution**

The batholiths are dominated by plutons of granodioritic-adamellitic compositions. There is a distinct bimodal separation of rocks of the mafic association from the dominant granodiorite-adamellite plutons. The compositional break occurs at the 55 - 65% SiO$_2$ range. In addition, the basic granitoid division into S- and I-type associations can be recognized in the East Coast Province. This fundamental tripartite division into mafic, I-type and S-type plutonic associations with only minor intermixing is important because it is believed that they represent mantle, lower crustal mafic granulitic and lower-middle crustal peraluminous metasedimentary sources respectively. This will be discussed further in a later section.

Unlike batholith terrains like the Coastal Batholith of Peru which consists of multiple suites intruded along the same lineament for over 70Ma (Pitcher, 1978), the East Coast Province is characterized by the emplacement of smaller batholithic masses (generally as 50-100 km strips) with a limited age range indicated (~10 - 20 Ma) for each batholith. The limited geochronological data indicate a progressive migration of batholith emplacement westwards towards the medial suture. Superimposed on this main westward younging trend is the intrusion of slightly younger, small, isolated plutons into older batholith belts. A single batholith can consist of various combinations and amounts of mafic plutons, mafic dykes, I-type suites and S-type suites. Such an intimate mingling of apparently unrelated magmas suggests that magmas were channeled along deep zones of weakness that facilitated their ascent and final emplacement (e.g. Pitcher, 1978).
(3) Granitoid differentiation processes

Chemical variation trends of the granitoid suites can be explained in terms of crystal-liquid fractionation equilibria and incomplete segregation of evolving silicic liquids and solid assemblages (liquidus crystals + restitic solids). Differentiation occurred in high-level magma chambers to produce the spectrum of granitoid rock-types. There is no need to postulate deep, hidden cumulate sequences. I-type liquid descent paths were controlled primarily by early plagioclase, hornblende and quartz fractionation whereas S-type liquid descent paths were controlled by early plagioclase, biotite and quartz fractionation. Major and trace element oxide modelling do not require extreme fractionation to achieve the chemical signatures displayed by the felsic granites, ~ 50 - 60% fractional crystallization is indicated although such figures are based on the assumption that crystal-liquid partition coefficients used are completely valid.

Liquid state differentiation processes are not required and if they operated, could only have caused minor superimposed effects on the strong crystal fractionation trends from granodioritic to granitic compositions. Certain patterns displayed by very evolved S-type granitic compositions (e.g. extreme depletion of Ba, Sr, Eu; LREE-depleted rare earth patterns and enrichment of U, Th, Rb) are similar to those of highly silicic rhyolite suites in which liquid differentiation processes are believed to be the dominant processes operating. Such patterns can also be explained by crystal fractionation. Conflicting behaviour of trace elements likely to form complexes i.e. depletion of Zr, Zn, Nb, La; constancy of Sn abundances and enrichment of Th, U in the S-types do not suggest an
important role for volatile complexing. Because highly differentiated S-type magmas will have high contents of \( \text{H}_2\text{O} \) and other volatiles, thus resulting in strongly depolymerized melts, the effect of melt structure control is not expected to be important. This is supported by the antithetic behaviour of the LREE and HREE and the depletion of Sr and Ba, characteristics that are opposed to those predicted by Ryerson and Hess's (1978) results for polymerized siliceous melts. It is suggested that liquid state differentiation processes are not important in controlling the chemical trends of the S-type suites except possibly in a handful of highly siliceous, highly evolved S-type granites whose chemical features may be the result of crystal-liquid fractionation equilibria, superimposed liquid differentiation processes and possibly, fluid phase interaction. There are no recognizable chemical patterns in highly evolved I-type granites that may be attributed to be the result of liquid state differentiation processes.

Fractionation-assimilation processes can be expected to be important in deeper parts of the crust (e.g. DePaolo, 1981a) but is not suspected to have significantly affected the high-level fractionation characteristics documented here. The trace element modelling however cannot differentiate between purely fractionation processes and combined assimilation-fractionation. The absence of any correlation between initial Sr and Nd ratios with \( \text{SiO}_2 \) contents together with mineral composition data (notably the biotite composition data) indicate that assimilation cannot be invoked to explain the generation of more siliceous I-type compositions from mafic parental compositions or be used to explain S-type granitoids as representing I-type magmas that have incorporated large amounts of sedimentary materials.
Magma sources

The tripartite division of the batholiths into mafic, I-type and S-type associations forms the main basis of the discussion in this section. Rocks of the mafic association can be plausibly attributed to be high-level fractionated components of mantle-derived primitive tholeiitic magmas. S-type suites can be explained as resulting from partial melting of metasedimentary sequences under amphibolite-granulite facies conditions involving muscovite and/or biotite breakdown reactions (e.g. Clemens and Wall, 1981). These highly peraluminous rocks are distinct from the peraluminous I-type adamellites and granites that can be produced by hornblende fractionation from metaluminous magmas. Such peraluminous I-type compositions are characterized by low-Al biotite, low to intermediate Sr initial ratios, high Ba (Figs. 17, 18) and low Th, U, Rb and Y. Early high-Al biotite, high Sr initial ratios and high Rb, Th, U, together with high alkali contents also negate the possibility that the peraluminous S-type magmas could result from removal of alkalis and related elements as soluble complexes from Al-saturated magmas.

The I-type batholiths are believed to be derived from partial melting of mafic to intermediate lower crust which may, in part, comprise of juvenile underplated materials. This mode of origin is implied by both the absence of rocks of intermediate compositions and the requirement of major and trace element oxide modelling that primitive melts be siliceous (~ 65 - 70% SiO₂) with LIL element abundances estimates in the region 70 - 120 ppm Rb, 250 - 350 ppm Sr and 350 - 700 ppm Ba. That siliceous melts can be generated by partial melting of basaltic rocks is well known and the data of, for example, Helz (1976) for melts in the 65 - 70% SiO₂ range are similar
to many of the observed compositions of the East Coast I-types.

Melting of mafic to intermediate (basaltic to andesitic bulk compositions) lower crust can therefore explain the bulk chemical compositions, low Sr initial ratios, low Th, U, Rb and the variable negative $\varepsilon_{\text{Nd}}$ values (as both old and juvenile underplated crust may be involved). A crude calculation of LIL element abundances of a melt derived from a hypothetical average lower crust with initial LIL element abundances estimated by Taylor (1979) assuming a granulite residue of plagioclase-hypersthene-clinopyroxene + quartz + hornblende yields figures of $\sim$50 ppm Rb, $\sim$400 ppm Sr and $\sim$450 ppm Ba. These are comparable to the abundances of primitive melt LIL element abundances estimated for the East Coast Province I-types. The model also is compatible with experimental observations that the liquidus mineralogy of tonalite-diorite magma compositions at lower crustal pressures are dominated by granulite mineralogies - plagioclase, pyroxene, quartz, garnet and hornblende (Wyllie et al., 1976). High temperatures required for partial melting of granulitic assemblages may be easily acquired by injection of basaltic magma into the lower crust. Cogent arguments and evidences for a mantle heat source to generate subduction-associated granitoid batholiths by partial melting of lower crustal rocks have been developed by Presnall and Bateman (1973), Hildreth (1981) and others.

Despite cautionary notes by Pitcher (1978) and Hildreth (1981), I have attempted to be explicit with regard to magma sources. This partly reflects the natural separation of the three plutonic associations (mafic, I-type and S-type) in the East Coast Province. The absence of widespread interaction between mafic and I-type magmas to produce intermediate suites and between I-types and S-types suggest melt generation under different conditions and different depths during
a single batholithic emplacement episode. A second reason for the explicit divisions is that the case of the East Coast Province may be indicating that the dominantly tonalitic compositions of many batholiths, which has led to suggestions that tonalites are primary magmas, may be the result of hybridization between mafic and dacitic magma compositions. One possible reason for the lack of hybridization between mafic and I-type magmas in the East Coast Province may be the short-lived nature of each batholithic emplacement episode ($\sim$10 - 20 Ma compared to $\sim$70 Ma for the Peruvian Coastal Batholith) after which a westward migration to another plutonic locus occurred. Thus, the absence of a long-lived plutonic locus discouraged the processes of continued mafic magma injection into a single source region, fractionation-assimilation, partial melting of host rocks and mixing of the variety of derived magmas.
CHAPTER SIX

GENESIS OF GRANITOID BATHOLITHS OF PENINSULAR MALAYSIA AND IMPLICATIONS FOR MODELS OF CRUSTAL EVOLUTION: EVIDENCE FROM A Nd-Sr ISOTOPIC AND U-Pb ZIRCON STUDY

INTRODUCTION

Voluminous granitoid batholiths found in linear plutonic belts constitute a distinctive feature of Phanerozoic fold belts. In view of their large volumes, the nature and origin of these batholiths are fundamental to our understanding of the generation and growth of continental crust. In the past, special attention has been given to the provenance of calcalkaline intrusive suites at convergent ocean-continent plate systems. Generalized models often invoke some mode of derivation of basaltic magmas from mantle sources with subsequent fractionation (+ crustal assimilation) to produce the gabbro-diorite-tonalite-granodiorite-granite plutonic association or the equivalent volcanic association. Such models suggest that continents grow via a process of marginal accretion of volcano-plutonic terrains continuously throughout the Phanerozoic. This seems compatible with the observed lateral addition of younger fold belts along the margins of old Precambrian shield areas. However, massive growth rates at the Archaean-Proterozoic boundary as suggested by some workers (Taylor, 1979; McCulloch and Wasserburg, 1978) and the observation that many continental margin batholiths invariably contain large amounts of recycled old continental crustal components (Allègre and Ben Othman, 1980) suggest that recent juvenile additions to the crust may not be too important.
In this study, Nd and Sr isotopic and U-Pb zircon data are presented for the Permian-Triassic batholiths of Peninsular Malaysia. The data are evaluated in a more general context by comparison with other batholith terrains and it will be shown that Phanerozoic granitoid batholiths that occur at plate margins seem to consistently contain components of mid-Proterozoic age. The implications of this observation will be examined.

Granitoids as images of continental crust

There is increasing awareness in recent years that granitoids can be used to provide images of underlying continental crust. Chappell and White (1974) and White et al. (1976) distinguished I-type granitoids (derived by melting of igneous source rocks) and S-type granitoids (derived by melting of sedimentary source rocks) in southeastern Australia and used the distribution of these granitoid types to define an 'I-S line' that seems to represent the most easterly extent of underlying metasedimentary Precambrian continental crust in southeastern Australia. Moore (1959), Kistler and Peterman (1973), DePaolo (1981b) and other American workers also recognized a Precambrian continent margin along western U.S.A. east of which initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of granitoids exceed 0.706 and granitoids display negative Nd values similar to those of old continental crust.

Isotopic tracer studies can lead to a better understanding of the compositional characteristics of granitoid source regions as well as giving first order age estimates of such source regions. Two isotopic methods allow age constraints to be made on granitoid source regions. These are the U-Pb zircon dating method and the Nd-Sm tracer method. Each method yields source chronological information via two totally independent approaches.
Inheritance in U-Pb zircon systems

The survival of old zircon components in zircon populations of young granitoids has been well documented by many studies (e.g. Grauert and Hofmann, 1973; Williams, 1977; Pidgeon and Aftalion, 1978). This phenomenon is widely termed zircon inheritance. The concordia curve (Wetherill, 1956), which is the locus of present day $^{205}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ratios for closed U-Pb systems of all ages, is widely used as a reference in discussions of inheritance patterns. On a concordia plot (Fig.1), a simple mixing between young magmatic zircon populations and old source zircons generates a chord (called a discordia chord) that intersects the concordia curve at two points and yields two intersection ages - a lower intersection age which, for the simple case considered here, represents the age of crystallization of the magmatic zircons and an upper intersection age which yields an estimate of the average age of the incorporated old zircon populations (case 1, Fig.1). Such discordia patterns which possess lower intersections that represent zircon crystallization ages are called reverse discordia. These differ from normal discordia which describe chords having upper intersections representing zircon crystallization ages (case 2, Fig.1). Normal discordia indicate absence of old zircon components and such discordia are generated by variable radiogenic Pb loss in different zircon fractions of a single population. For the case where young zircon populations contain inherited components and have also suffered Pb loss (case 3, Fig.1), complex non-linear patterns may be produced which are difficult to interpret.
Fig. 1 Discordant U-Pb systems on a concordia diagram. In most cases, the two U-Pb apparent ages obtained for a single zircon fraction do not agree i.e. the point falls off concordia and is described as being discordant. Crystallization ages are obtained by extrapolating a discordance line joining a number of such discordant points to the point of either (1) zero inherited zircon component (case 1 in Fig.) or (2) zero Pb loss (case 2 in Fig). Case 3 represents the situation when both Pb loss and inheritance have both affected a zircon population.
The presence of inherited zircons in young granitoids is one evidence that old crustal components have been incorporated in young granitoid magmas. Where Pb loss is not important, zircon inheritance ages allow an estimate of the ages of the zircon-bearing protoliths to be made. For S-type granitoids, inheritance ages are estimates of the ages of the pre-sedimentary stage protoliths. Well studied terrains with a consistent age range yielded by inherited zircon components from unrelated suites lend support to the belief that such age estimates represent those of the underlying crystalline basement that are the source regions for granitoid magmas.

Nd model ages and a proposed model depleted mantle evolution

The Sm-Nd system has been used recently by a number of workers to study the petrogenesis of granitoids (Allègre and Ben Othman, 1980; Hamilton et al., 1980; DePaolo, 1981b; McCulloch and Chappell, 1982). The importance of old crustal components in Phanerozoic granitoids was shown by the results of these studies.

The important principles behind the use of the Sm-Nd tracer approach are the relative immobility of the rare earth elements (REE) and difficulty in fractionating Sm and Nd by normal crustal processes. As such, Nd isotopes have proved useful for estimating the times when crustal materials were first extracted from the mantle. The approach is based on calculation of model ages assuming that Sm and Nd in a rock remain unFractionated from the time it was initially derived from a mantle reservoir. Early models (e.g. McCulloch and Wasserburg, 1978; DePaolo and Wasserburg, 1979) calculated 'mantle extraction ages' by assuming that crustal materials were derived from a chondritic mantle reservoir. Such ages are termed $T_{\text{Nd}}^{\text{CHUR}}$ ages. The expression for $T_{\text{Nd}}^{\text{CHUR}}$ is derived from the following relationships:
The present day $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of a sample is related by

$\alpha$-decay of $^{147}\text{Sm}$ through the decay equation:

$$
(143\text{Nd}/144\text{Nd})_{\text{sample}} = (143\text{Nd}/144\text{Nd})_{\text{initial}} + (147\text{Sm}/144\text{Nd})_t (e^{-\lambda t} - 1) \tag{1}
$$

A similar expression can be written for a CHUR mantle,

$$
(143\text{Nd}/144\text{Nd})_{\text{CHUR}}^0 = (143\text{Nd}/144\text{Nd})_{\text{CHUR}}^t + (147\text{Sm}/144\text{Nd})_{\text{CHUR}} (e^{-\lambda t} - 1) \tag{2}
$$

For the time $T$ when a rock is fractionated from a CHUR mantle,

$$
(143\text{Nd}/144\text{Nd})_{\text{initial}}^\text{CHUR} = (143\text{Nd}/144\text{Nd})_{\text{CHUR}}^t . \text{ Substituting}

(143\text{Nd}/144\text{Nd})_{\text{CHUR}}^t = (143\text{Nd}/144\text{Nd})_{\text{CHUR}}^o - (147\text{Sm}/144\text{Nd})_{\text{CHUR}} (e^{\lambda t} - 1) \text{ into equation (1) yields :}

$$
\ln \frac{1}{\lambda} \ln \left[ 1 + \frac{(143\text{Nd}/144\text{Nd})_{\text{meas}} - (143\text{Nd}/144\text{Nd})_{\text{CHUR}}^o}{(147\text{Sm}/144\text{Nd})_{\text{meas}} - (147\text{Sm}/144\text{Nd})_{\text{CHUR}}} \right] \tag{3}
$$

where $(143\text{Nd}/144\text{Nd})_{\text{CHUR}}^o = 0.511836$ and $(147\text{Sm}/144\text{Nd})_{\text{CHUR}} = 0.1967$ .

As more data became available, increasing evidence was found for the involvement of depleted mantle (i.e. mantle having a higher Sm/Nd ratio compared to chondrites) in the generation of continental crust. This evidence includes:

1. The occurrence of I-type granitoid plutons possessing depleted mantle isotopic signatures e.g. Sierra Nevada and Peninsular Ranges Batholiths of California (DePaolo, 1981b); Bega and New England Batholiths of southeastern Australia (McCulloch et al., 1982). Such plutons may be associated with plutons having less depleted signatures that are often interpreted to have resulted from assimilation of continental crust.
2. The widespread distribution of modern-day island arc and continent margin basalt-andesite-dacite volcanic series having depleted mantle isotopic characteristics and which are often related in space and time to granitoid suites (McCulloch and Perfit, 1981).

3. Results from terrains like the Colorado Front Range where DePaolo (1981c) showed that a crustal segment originally derived from a depleted mantle source could be tapped by subsequent magmatic episodes without involving further additions from the mantle.

4. Depleted mantle model ages, $T_{\text{DM}}^{\text{Nd}}$, are more comparable to crystallization ages when mantle-derived materials of Proterozoic ages are studied (DePaolo, 1981c).

5. The occurrence of mantle-derived rocks throughout the Archaean and Proterozoic with positive $\epsilon_{\text{Nd}}$ initial ratios (DePaolo, 1981c; McCulloch and Compston, 1981; Duyverman et al., 1982).

Such observations show that, in many terrains, juvenile crust was derived not from a chondritic mantle but a mantle previously depleted in incompatible elements. Depleted mantle model ages ($T_{\text{DM}}^{\text{Nd}}$) are consequently regarded as more realistic estimates of mantle extraction ages than $T_{\text{CHUR}}^{\text{Nd}}$ ages in this study. $T_{\text{DM}}^{\text{Nd}}$ ages are obtained by substituting depleted mantle parameters into equation (3), yielding:

$$T_{\text{DM}}^{\text{Nd}} = \frac{1}{\lambda_{\text{Sm}}} \ln \left[ 1 - \frac{\frac{(143\text{Nd}/144\text{Nd})_{\text{meas}}}{(147\text{Sm}/144\text{Nd})_{\text{meas}}} - \frac{(143\text{Nd}/144\text{Nd})_{\text{DM}}}{(147\text{Sm}/144\text{Nd})_{\text{DM}}}}{0.512348} \right]$$

where $(143\text{Nd}/144\text{Nd})_{\text{DM}} = 0.512348$ and $147\text{Sm}/144\text{Nd} = 0.225$.

The model depleted mantle evolution used in this study is based on the assumption of LREE depletion from a primitive chondritic mantle at 2700 Ma and that the present day $143\text{Nd}/144\text{Nd}$ value of the depleted mantle reservoir is 0.512348 (i.e. $\epsilon_{\text{Nd}}(0) = +10$). These
parameters were chosen based on the following reasons. Evidence for the widely held view that the period between 2500-3000Ma represents a major episode of production of juvenile continental crust (McCulloch and Wasserburg, 1978; Taylor, 1979) are compelling on geological, isotopic and geochemical grounds. If this is correct, significant volumes of a complementary depleted mantle reservoir must be produced near the Archaean-Proterozoic boundary. An intermediate age of 2700 Ma was selected but it makes little difference to calculated $T_{\text{DM}}^{\text{Nd}}$ ages whether 2500 Ma or 3000 Ma is chosen as the time of initiation of the depleted mantle evolution. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.512348 was chosen because it is the average value of MORB, the most widespread contemporary depleted mantle reservoir with unambiguous and well-defined characteristics. Fig.2 shows the evolution line of the proposed model depleted mantle and compares its relationships to the retarded $^{143}\text{Nd}/^{144}\text{Nd}$ evolution of primitive chondritic mantle and continental crust which both possess higher $\text{Sm/Nd}$ ratios. The proposed depleted mantle evolution is, however, regarded only as a reasonable approximation suitable for depleted mantle model age calculations. More data are required to define a complex mantle evolutionary history. Heterogeneities can exist in the Archaean mantle, as documented by McCulloch and Compston (1981).

It is stressed that the $T_{\text{DM}}^{\text{Nd}}$ ages calculated as suggested above represent model ages that assume derivation from a model depleted mantle. One aim in proposing the present model depleted mantle evolution was to facilitate more widespread use of $T_{\text{DM}}^{\text{Nd}}$ ages. As such, only a simple single stage evolution path has been discussed. Models describing the Nd isotopic evolution of a 'continuously depleting' upper mantle have been proposed by DePaolo (1981c) and Allegre (1982). Their use complicates computation of $T_{\text{DM}}^{\text{Nd}}$ ages but
Fig. 2. Evolution of $^{143}\text{Nd}/^{144}\text{Nd}$ (expressed as units) with time for several model reservoirs. In this figure, the evolution of a depleted mantle reservoir and continental crust are conveniently referenced to that of a chondritic mantle which yields a horizontal line with $\varepsilon_{\text{Nd}} = 0$. Continent crustal rocks with LREE-enriched patterns (i.e., $^{147}\text{Sm} / ^{144}\text{Nd} < \text{chondrite}$) yield a retarded $^{143}\text{Nd}/^{144}\text{Nd}$ evolution compared to chondrites. As such, continental crust yields increasingly negative $\varepsilon_{\text{Nd}}$ with time compared to a chondritic mantle. In contrast, enhanced evolution of $^{143}\text{Nd}/^{144}\text{Nd}$ due to high Sm/Nd for a depleted mantle leads to increasingly positive $\varepsilon_{\text{Nd}}$ values with time. Intersections of the arrows representing continent crustal evolution lines with the CHUR evolution line gives $T_{\text{CHUR}}^{\text{Nd}}$ ages whereas intersections with the depleted mantle evolution line yields $T_{\text{DM}}^{\text{Nd}}$ ages. For model ages $> 2.0$ Ga the difference between $T_{\text{CHUR}}^{\text{Nd}}$ and $T_{\text{DM}}^{\text{Nd}}$ ages is insignificant.
yield ages not significantly different from those calculated using the present model.

ε -notation

The ε-notation for Nd and Sr is now widely used to describe variations of initial $^{143}$Nd/$^{144}$Nd and initial $^{87}$Sr/$^{86}$Sr ratios compared to a primitive chondritic mantle. Initial $^{143}$Nd/$^{144}$Nd ratios are expressed as $\varepsilon_{Nd}(t)$ using the expression:

$$\varepsilon_{Nd}(t) = \left[ \frac{(143 \text{Nd}/144 \text{Nd})_{\text{sample}}^t}{(143 \text{Nd}/144 \text{Nd})_{\text{CHUR}}^t} - 1 \right] \cdot 10^4 \quad (5)$$

Initial ratios are calculated from present day measurements by the conventional radioactive decay equation:

$$\left( \frac{143 \text{Nd}/144 \text{Nd}}{147 \text{Sm}/144 \text{Nd}} \right)_{\text{meas}} = \left( \frac{143 \text{Nd}/144 \text{Nd}}{147 \text{Sm}/144 \text{Nd}} \right)_{\text{initial}} + \left( \frac{147 \text{Sm}/144 \text{Nd}}{147 \text{Sm}/144 \text{Nd}} \right)_{\text{meas}} (e^{\lambda t} - 1) \quad (6)$$

$$(143 \text{Nd}/144 \text{Nd})_{\text{CHUR}}^t$$ can be calculated also from equation (6) for any age by substituting present day $143 \text{Nd}/144 \text{Nd} = 0.511836$ and $147 \text{Sm}/144 \text{Nd} = 0.1967$.

An identical notation can be derived for the Sr system i.e.

$$\varepsilon_{Sr}(t) = \left[ \frac{(87 \text{Sr}/86 \text{Sr})_{\text{sample}}^t}{(87 \text{Sr}/86 \text{Sr})_{\text{CHUR}}^t} - 1 \right] \cdot 10^4 \quad (7)$$

CHUR parameters adopted in this study for Sr are $87 \text{Sr}/86 \text{Sr} = 0.7048$ (present day) and $87 \text{Rb}/86 \text{Sr} = 0.085$. Derivation of initial ratios is analogous to that for the Nd system using equation (6).
Nd MODEL AGES AND U-Pb ZIRCON INHERITANCE AGES

Under ideal circumstances, zircon inheritance ages should yield the age of an ultimate granitoid or high-grade metamorphic basement source whereas Nd model ages should approximate mantle extraction ages. Juvenile additions to the continental crust, whether laterally accreted or underplated, may be subsequently subjected to one or more partial melting episodes to form granitoid batholiths. Granitoids should therefore possess Nd model ages ≥ zircon inheritance ages. In practice, data sets are invariably more complicated. Pb loss phenomena in old source zircons would yield younger inheritance ages and Nd model ages are dependent on the choice of $T^{\text{Nd}}_{\text{DM}}$ or $T^{\text{Nd}}_{\text{CHUR}}$ ages. In addition, complex mantle-crust interactions may occur at settings where continental margin I-type magmas are generated. Assimilation of old crustal zircons by mantle-derived magmas could conceivably result in I-type magmas having zircon inheritance ages older than calculated mantle extraction ages. However, the large degree of assimilation required to stabilize significant zircon inheritance effects would in general result in the crustal Nd signature dominating the depleted mantle contributions to the hybrid magma. This is because even wholly crustal-derived S-type granitoid magmas display only limited zircon inheritance effects. Bearing in mind such theoretical limitations and the fact that the discussion is only concerned with broad age ranges, the zircon inheritance and $T^{\text{Nd}}_{\text{DM}}$ age data for the Peninsular Malaysian batholiths will be discussed and compared with that of other batholith terrains. The implications of the data for crustal growth models will then be examined.
The use of the Rb-Sr system to estimate model ages has been omitted in view of the strong fractionation of these elements during igneous differentiation and their high mobility in sedimentary and metamorphic environments.

EXPERIMENTAL TECHNIQUES

Sm-Nd

Experimental procedures for Sm-Nd analyses are those of McCulloch and Chappell (1982). Samples spiked with $^{150}\text{Nd}$ and $^{147}\text{Sm}$ tracers were dissolved in Teflon bombs with HF-HClO$_4$ and then converted to chloride solution form. Next, REE were separated from major elements using conventional cation exchange resins. Sm and Nd were then separated on a second column using 0.2M 2-methylactic acid with pH = 4.6. Nd was measured as Nd$^+$ on a modified MSZ mass spectrometer using Re triple filaments. Possible interferences caused by Ce and Sm were checked before runs at masses 140 and 152 respectively. In all cases interferences were negligible. $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are normalised to $^{146}\text{Nd}/^{142}\text{Nd} = 0.636151$. $^{143}\text{Nd}/^{144}\text{Nd}$ obtained for BCR-1 in this laboratory is 0.511843 ± 20.

Rb-Sr

Sr measurements were made on the identical samples processed for Sm-Nd work. Sr was loaded on single Ta filaments using HCl - $\text{H}_3\text{PO}_4$. Measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. At the high current settings for these Sr runs, Rb interference at mass 87 is insignificant. $^{87}\text{Sr}/^{86}\text{Sr}$ measurements for NBS 987 on the MSZ mass spectrometer determined during this period average 0.71029 ± 3. Rb was obtained mainly by XRF techniques discussed in Chappell et al. (1969) but some samples were
also determined by isotope dilution. Results agree to within \( \pm 2\% \).

**U-Pb**

Techniques of zircon chemistry and mass spectrometry follow those of Page (1981). Zircons, separated by conventional Wifley table, heavy liquid and magnetic methods, were handpicked into individual size fractions for analyses. Chemical dissolution techniques adopted are those of Krogh (1973). Pb was analysed by the silica-gel technique and U was loaded using Ta_2O_5 and H_3PO_4 on single Re filaments and analysed as UO\(^+\). Uncertainties of \( \pm 0.3\% \) and \( \pm 1.0\% \) are estimated for \( ^{207}\text{Pb} / ^{206}\text{Pb} \) and Pb/U respectively. Where applicable, discordia chords have been calculated by the method of least squares, using a correlation coefficient of 0.96. Intersections with condordia were determined by allowing the slope of the chord to vary about its centroid by 2\( \sigma \) limits. The total Pb blank is typically 0.3 ng per analysis.

**EAST AND WEST COAST GRANITOID PROVINCES OF PENINSULAR MALAYSIA**

The detailed petrographic and chemical characteristics of the Peninsular Malaysian batholiths were discussed at length in chapters four and five, so only arguments necessary to develop a discussion of the isotopic work are outlined in this chapter. To describe the batholiths, the terminology of S- and I-type granites as proposed by Chappell and White (1974) for the Palaeozoic granitoid batholiths in the Lachlan Fold Belt of southeastern Australia has been adopted. Thus, S-types are believed to be derived from old metasedimentary crust that has undergone considerable weathering and recycling processes whereas I-types are derived from igneous or metaigneous materials that may or may not be directly mantle-derived but which
have not undergone significant weathering. The usage, however, does not imply exact correspondence to the chemical definitions outlined by Chappell and White (1974) as there are important major and trace element differences between Malaysian and southeastern Australian granitoids.

Batholiths east of the medial collision suture in Peninsular Malaysia are mainly exposed as a series of north-south trending masses distributed along a narrow eastern coastal belt (Fig.3). Bignell and Snelling (1977) termed this belt the East Coast Province. The batholiths of the East Coast Province are intrusive into Carboniferous to Triassic sedimentary formations which contain abundant interbedded basalt-andesite-dacite-rhyolite volcanic sequences that are comparable in age and compositional range to the granitoid batholiths.

The most common granitoid-type in this East Coast Province is a biotite + hornblende granodiorite, although individual intrusive complexes may comprise of rocks ranging from gabbro to potassic granite. The calcalkaline affinities of the East Coast batholiths are displayed on an AFM diagram (Fig.4) and their metaluminous I-type characteristics are shown in Fig.5. A volumetrically minor group of peraluminous granitoid plutons having S-type affinities (high K, Rb, Th, U, Pb and high Sr initial ratios) is closely associated spatially and temporarally with the dominant I-type plutons.

The Triassic granitoid batholiths of the West Coast Province found west of the collision suture are exclusively of an S-type nature. The most common rock-type is a coarse-grained, leucocratic biotite adamellite with large, coarsely perthitic K-feldspar megacrysts. There is a rather restricted compositional range of rock-types from granodiorite to granite. Aplitic and pegmatitic
Fig. 3. Map showing generalized distribution of granitoid batholiths in Peninsular Malaysia and sample localities. Age distribution is based on the interpretation of data of Bignell and Snelling (1977) and U-Pb zircon and Rb-Sr results discussed in chapter three.
Fig. 4. Chemical trends for granitoids of the West Coast Province and East Coast Province (including associated mafic complexes) on AFM diagrams. Includes data of Kumar (1981).
Fig. 5. Fields for the West Coast and East Coast Province granitoids on modified ACF triangular diagrams.
The peraluminous mineral phases present are high Al-biotite, muscovite, tourmaline and rare Mn-rich almandine garnet. Notable chemical characteristics include high to very high abundances of K, Rb, Th, U, Pb and high Sr initial ratios.

The tectonic settings of the Peninsular Malaysian granitoid batholiths were discussed in chapters two, four and five and were based on the tectonic reconstruction of Mitchell (1977). Mitchell considered the predominantly I-type granitoid batholiths of the East Coast Province to represent a Permian to Triassic plutonic arc formed at a continental margin subduction setting (Fig. 6A). In contrast, the late Triassic granitoid batholiths of the West Coast Province were believed to have formed in a continent collision setting (Fig. 6B) that developed subsequent to the closure of an ocean/marginal sea by subduction.

S-TYPE BATHOLITHS OF THE WEST COAST PROVINCE

Geochronology

The interpretation of the geochronology of the batholiths is based on the results of the Rb-Sr and K-Ar work of Bignell and Snelling (1977) and the U-Pb zircon and Rb-Sr work discussed in chapter three.

A striking feature of the radiometric results from the West Coast Province is the discordant nature of K-Ar ages compared to U-Pb zircon and Rb-Sr whole rock isochron ages. K-Ar ages range from 40 to 210 Ma whereas U-Pb zircon and Rb-Sr isochrons record late Triassic (200 - 220 Ma) events. The evidence for widespread Ar loss is overwhelming and the variable loss patterns have been attributed to major fault-related disturbances (Bignell and Snelling, 1977). U-Pb zircon
Fig. 6A-B Highly schematic plate tectonic reconstruction illustrating the settings in which the Peninsular Malaysian granitoid batholiths were formed (after Mitchell, 1977).
Dating yields precise and least ambiguous estimates of emplacement ages (198 to 219 Ma). Zircon dating of several plutons which possessed problematic Rb-Sr systematics that led to suggestions of possible Palaeozoic events have shown that these plutons are late Triassic in age and characterized by high (0.720 - 0.730) and variable Sr initial ratios. The emplacement ages recorded by both U-Pb zircon and Rb-Sr whole rock systems are interpreted to represent the time of the postulated late Triassic continental collision event.

U-Pb zircon systems and zircon inheritance

Results of zircon analyses from nine localities (Table 3) in the exclusively S-type West Coast Province may be conveniently divided into 'concordant' and distinctly discordant groups. The 'concordant' group plotted in Fig. 7A yield ages of 198 - 215 Ma. Some samples approach 'ideal concordance' in that separate zircon fractions have $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages agreeing to within 1 Ma. Because of the near-linear nature of the concordia in the working age range, $^{207}\text{Pb}/^{206}\text{Pb}$ ages are too variable and are of little use.

The discordant data (Fig. 7B) display short reverse discordance patterns (<10% of discordia chord) that extrapolate to intersect concordia giving lower intersection ages of 211 - 218 Ma. This age range falls within that for the 'concordant' group. The geological consistency and concordance of much of the U-Pb data suggest insignificant post-emplacement Pb-loss. Consequently, the upper intersection ages of the reverse discordia, consistently in the 1500 - 1700 Ma range (except for sample 45 having a 1250 Ma upper intersection) may broadly represent the ages of the old zircon components - assuming the simple interpretation that reverse discordia are undisturbed mixes of old source zircons and overgrowths of young
<table>
<thead>
<tr>
<th>Sample</th>
<th>Upper Intersection (Ma)</th>
<th>Lower Intersection (Ma)</th>
<th>Emplacement age indicated by U-Pb zircon system</th>
</tr>
</thead>
<tbody>
<tr>
<td>13</td>
<td>1560 ± 150</td>
<td>215 +5 -6</td>
<td>Lower intersection of 215±6 Ma</td>
</tr>
<tr>
<td>14</td>
<td>1700 ± 600</td>
<td>209 ± 2</td>
<td>Lower intersection of 209±2 Ma</td>
</tr>
<tr>
<td>29</td>
<td>1650 ± 400</td>
<td>215 +2 -4</td>
<td>Lower intersection of 215±4 Ma. Zircons from both samples belong to cogenetic suite and display collinear relationships.</td>
</tr>
<tr>
<td>39</td>
<td>1610 ± 450</td>
<td>211 +5 -8</td>
<td>Lower intersection of 211±5 Ma.</td>
</tr>
<tr>
<td>38</td>
<td>-</td>
<td>-</td>
<td>Three concordant fractions give 198 ± 2 Ma.</td>
</tr>
<tr>
<td>41</td>
<td>-</td>
<td>-</td>
<td>Three concordant fractions give 206 ± 2 Ma.</td>
</tr>
<tr>
<td>45</td>
<td>1240 ± 300</td>
<td>208 ± 2</td>
<td>Lower intersection of 208±2 Ma.</td>
</tr>
<tr>
<td>78</td>
<td>1550 ± 350</td>
<td>219 +5 -9</td>
<td>Lower intersection of 219±5 Ma.</td>
</tr>
</tbody>
</table>
Fig. 7A-B. Summary of concordant and discordant zircon data for the West Coast Province plotted on concordia diagrams.
magmatic zircons. The old components present in the zircons are
evidence that old Precambrian components have been incorporated in the
late Triassic granitoids of the West Coast Province. Although the old
source zircons may have suffered Pb loss before being incorporated
into the magmas, in which case the upper intersections underestimate
the true ages of the zircons, the consistency of the upper
intersections is suggestive that only minor Pb loss occurred. In the
latter case, the upper intersections should provide reliable estimates
of the ages of the old zircon populations.

\[ \varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}} \]

Eight samples from representative granitoid suites, two volcanic
samples and two metasedimentary samples from the West Coast Province
were studied using the Sm-Nd and Rb-Sr methods. The analytical data
are shown in Tables 2 and 3. Initial ratios are calculated using
intrusive ages based on evaluation of the U-Pb zircon, Rb-Sr whole
rock isochron and K-Ar mica ages discussed in chapter three. \( \varepsilon_{\text{Nd}} \)
(T) values for the metasediments were age corrected to 200 Ma so as to
facilitate a direct comparison with the granitoid data.

Samples from the S-type batholiths of the West Coast Province
define a subhorizontal array on an \( \varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}} \) diagram, (Fig.8) with
\( \varepsilon_{\text{Nd}} \) values of -6 to -10 and very high \( \varepsilon_{\text{Sr}} \) values of +160 (0.716)
to +600 (0.748). This subhorizontal array results from the
preferential retention of Rb relative to Sr in clays during
sedimentary weathering cycles, resulting in high Rb/Sr ratios in the
metasedimentary source rocks and consequently, enhanced evolution of
\( ^{87}\text{Sr}/^{86}\text{Sr} \). In contrast to the large \( \varepsilon_{\text{Sr}} \) range, the relatively
restricted \( \varepsilon_{\text{Nd}} \) range suggests that the Sm-Nd system has not been
substantially fractionated by sedimentary and metamorphic processes
<table>
<thead>
<tr>
<th>Sm</th>
<th>Nd</th>
<th>Rb</th>
<th>Sr</th>
<th>87Rb/86Sr</th>
<th>143Sm/144Nd</th>
<th>87Sr/86Sr</th>
<th>147Nd/144Nd</th>
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<tr>
<td>13</td>
<td>7.82</td>
<td>36.93</td>
<td>409</td>
<td>62</td>
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<td>74</td>
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<td>11.27</td>
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<td>122</td>
<td>9.38</td>
<td>53.82</td>
<td>260</td>
<td>212</td>
<td>3.55</td>
<td>0.1055</td>
<td>0.73609</td>
</tr>
</tbody>
</table>

Meta-sediments

| 83 | 4.71| 25.47| 81 | 184 | 1.27 | 0.1119 | 0.72558 | 0.51138 ± 0.4 |
| KH-1 | 3.35| 17.28| 92 | 36 | 7.00 | 0.1174 | 0.75943 | 0.51117 ± 0.4 |

* denotes duplicate sample
<table>
<thead>
<tr>
<th>Age (Ma)</th>
<th>$^{143}\text{Nd}$ (I)</th>
<th>$^{87}\text{Sr}$ (I)</th>
<th>$\varepsilon_{\text{Nd}}$ (I)</th>
<th>$\varepsilon_{\text{Sr}}$ (I)</th>
<th>$^{\text{Nd}}_{\text{CHUR}}$ (Ma)</th>
<th>$^{\text{Nd}}_{\text{DM}}$ (Ma)</th>
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<td>13</td>
<td>210</td>
<td>0.51106</td>
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<td>-9.8</td>
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<td>0.7241</td>
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<td>1080</td>
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<td>0.7159</td>
<td>-8.1</td>
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<td>1000</td>
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<td><strong>Meta</strong></td>
<td><strong>sediments</strong></td>
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<td>****</td>
<td>****</td>
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<td>****</td>
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<td>0.7395</td>
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</table>
Fig. 8. Plot of initial $\varepsilon_{\text{Nd}}$ versus $\varepsilon_{\text{Sr}}$ for the West Coast Province granitoid data listed in table 3. $T_{\text{DM}}^{\text{Nd}}$ ages shown on right-hand vertical axis represents model ages calculated by using a typical continental crust $\frac{^{147}\text{Sm}}{^{144}\text{Nd}}$ ratio of 0.12 in equation (4). Mantle array after DePaolo and Wasserburg (1979).
and has retained a memory of the presedimentary cycle source rocks that is potentially of invaluable aid in studying granitoid sources. Because sedimentary processes tend to average or homogenize originally differing provenance components, the Nd isotopic signatures of sedimentary rocks tend to provide bulk images of the exposed continental crust feeding the sedimentary basin. In relation to S-type granitoids, the contrasting effects of sedimentary processes on the Sm-Nd and Rb-Sr system result in greatly increased $\varepsilon_{\text{Sr}}$ values along the horizontal axis and a contraction of the variation of $\varepsilon_{\text{Nd}}$ values along the vertical axis, thus resulting in the characteristic subhorizontal patterns often displayed by S-type batholiths.

$^{147}\text{Sm}/^{144}\text{Nd}$ ratios of the samples retain typical granitoid values of 0.10 to 0.13 suggesting Nd and Sm have not been fractionated significantly by the igneous differentiation processes. On the other hand, extreme fractionation of Rb and Sr can be observed in some of the felsic plutons which may possess very high $^{87}\text{Rb}/^{86}\text{Sr}$ ratios exceeding 60. Extensive plagioclase fractionation best explains this chemical feature.

$^{\text{T}_{\text{Nd}}}_{\text{DM}}$ ages, zircon inheritance ages and basement age implications

Ages calculated using equation (4) are listed in Table 3. These ages range from 1300 to 1800 Ma. They are consequently in general agreement with U-Pb zircon upper intersection ages of 1500 to 1700 Ma (with the exception of one 1250 Ma intersection). Two qualifications however need to be stated. Firstly, the long extrapolation to upper intersections with the concordia curve results in errors in the order of ± 300 Ma for the individual intersection ages. Secondly, the Nd model ages have been calculated assuming derivation from depleted mantle. For reasons discussed earlier in
this chapter, evidence for the involvement of depleted mantle in post-Archaean continental crust generation is compelling. If this basic assumption is correct, then agreement obtained from the two methods cannot be coincidental even if the errors in age estimates may be large. The discussion is however only concerned with broad age ranges. The agreement between U-Pb zircon inheritance ages and $T_{DM}^{Nd}$ ages is within the resolution required for the present purposes. Such agreement indicates that extraction of juvenile crust from the mantle and formation of zircon-bearing Proterozoic granitoid protoliths represented a single event or events that were not separated by an extended time interval (100-200 Ma). Two samples of the Palaeozoic metasedimentary rocks of the West Coast Province possess similar $\varepsilon_{Nd}$ values (age corrected to 200 Ma) and $T_{DM}^{Nd}$ ages (1300 Ma and 1660 Ma) as the granitoids. This is consistent with their derivation from the same cratonic source as the metasedimentary source rocks for the S-type granitoid batholiths. On the basis of the available data, the original 'crust formation' age of the West Coast Province continental segment is estimated to be ~1500 - 1700 Ma, the age overlap obtained from both the U-Pb and Sm-Nd systems. Although basement of this age has not previously been recognised in Peninsular Malaysia, the data together with stratigraphic evidence (Stauffer, 1974) supports the argument for an underlying Precambrian basement. In addition the isotopic results indicate a mid-Proterozoic basement age. It is suggested that the 'missing' western craton postulated by Stauffer (1974), or at least its detachment margin, is also of this age.
GRANITOID BATHOLITHS OF THE EAST COAST PROVINCE

Geochronology

K-Ar cooling ages, together with more limited Rb-Sr whole rock isochron ages and U-Pb zircon ages indicate emplacement ages for the granitoid batholiths of the East Coast Province as Permian to Triassic. These are consistent with broad stratigraphic constraints (Chapters three and five). Bignell and Snelling (1977) showed that K-Ar ages are generally concordant with Rb-Sr whole rock isochron ages. The limited U-Pb zircon data discussed in chapter three are consistent with this contention. Ages show a general younging trend towards the medial suture and suggest a westward migration of plutonic emplacement. Two intrusive episodes are indicated by the radiometric results for the East Coast Province - early Permian (255 - 270 Ma) and late Permian-early Triassic (220 - 240 Ma). A late Triassic event was documented by Bignell and Snelling for small granitoid masses in the adjacent Central Province (Fig.4).

U-Pb systems and zircon inheritance

U-Pb zircon work was carried out on three plutons crudely representing the extremes of the broad spectrum of granitoid types.

Sample 132 (see locality map, Fig.4) is an I-type biotite-hornblende granodiorite pluton intrusive into rocks of Carboniferous age. A K-Ar biotite age of 264 + 7 Ma was reported by Bignell and Snelling (1977). The four zircon fractions analysed for this sample define a reverse discordia (Fig.9) with a lower intersection age of 257+ Ma which agrees with the K-Ar age within analytical error. The upper intersection age is ~ 800 Ma i.e. the presence of old inherited zircons is indicated.
Fig. 9. Summary of U-Pb zircon data for the East Coast Province on a concordia diagram. Ages located beside arrows denote upper intersection ages with concordia.
Sample 120 is a felsic I-type biotite adamellite. A three-point Rb-Sr isochron yields an age of $220 \pm 5$ Ma (chapter three). The U-Pb zircon data for this sample define a normal discordia indicating that old zircons have not been incorporated by this rock. The discordia chord has an upper intersection age of $227 \pm 2$ Ma. Although the U-Pb zircon and Rb-Sr isochron age lie just outside of the estimated analytical uncertainties, a lower Triassic emplacement age is clearly indicated.

Sample 106 is from an S-type adamellite stock that gave a $252 \pm 7$ Ma Rb-Sr whole rock isochron age and a $217 \pm 8$ Ma K-Ar biotite age (chapter three). A single zircon fraction analysed falls within analytical error on concordia to give an age of $264 \pm 2$ Ma. This is slightly older than, but supports the Permian Rb-Sr isochron age. Two other fractions are discordant and together with the concordant fraction yield a reverse discordia chord with an upper intersection of $\sim 1350$ Ma.

For all three plutons studied, the U-Pb zircon ages are compatible with field relationships and either less reliable Rb-Sr isochron ages or biotite K-Ar ages. In common with many S-type plutons, sample 106 displays inherited zircon phenomena. The I-type plutons show more variable behaviour - zircons from sample 132 contain inherited components whereas zircons from sample 120 do not. The reverse discordia upper intersections of $\sim 800$ Ma and $\sim 1350$ Ma for sample 132 and sample 106 respectively indicate the presence of Precambrian components in these Permian - Triassic granitoids and suggests that a Precambrian basement underlies the East Coast Province.
Eight samples of the I-type association and five samples of the S-type association in the East Coast Province were analysed for their Nd and Sr isotopic compositions. The results are displayed in Tables 4 and 5. Age corrections represent best estimates of emplacement ages based on the accumulated U-Pb zircon, Rb-Sr and K-Ar data discussed in chapters three and five. Initial $\varepsilon_{\text{Nd}}$ vs. $\varepsilon_{\text{Sr}}$ values are shown in Fig. 10.

Initial $\varepsilon_{\text{Nd}}$ and $\varepsilon_{\text{Sr}}$ values for the I-type granitoids range from -0.8 to -6.2 and +10 (0.705) to +80 (0.710). The field for the I-type samples overlaps completely with the field for rocks of the mafic (generally gabbroic) association of the East Coast Province (chapter five). The contrasting chemistries between the I-type plutons and the mafic rocks and the bimodal character of the province indicated by the scarcity of intermediate compositions (i.e. diorites, quartz-diorites, monzodiorites), however, argue against any simple genetic relationship between I-type plutons and the mafic rocks. In addition, the isotopic data are not consistent with a model requiring mantle-derived mafic magmas with depleted characteristics (positive initial $\varepsilon_{\text{Nd}}$, negative initial $\varepsilon_{\text{Sr}}$) to have fractionated and assimilated continental crust resulting in intermediate to felsic granitoids displaying 'crustal' (negative initial $\varepsilon_{\text{Nd}}$, positive initial $\varepsilon_{\text{Sr}}$) signatures. The East Coast Province data are more compatible with separate multi-component source models - enriched sub-continental mantle sources for the mafic magmas but derivation of the I-type plutons by partial melting of old lower crust and juvenile underplated crust.
<table>
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* duplicate sample
**TABLE 5**

Nd and Sr isotopic characteristics of the East Coast Province granitoids

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<th>Age (Ma)</th>
<th>( \frac{^{143}Nd}{^{144}Nd} )</th>
<th>( \frac{^{87}Sr}{^{86}Sr} )</th>
<th>( \varepsilon_{Nd}^{(1)} )</th>
<th>( \varepsilon_{Sr}^{(1)} )</th>
<th>( T_{CHUR}^{Nd} ) (Ma)</th>
<th>( T_{DM}^{Nd} ) (Ma)</th>
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A \(^{147}Sm/^{144}Nd \approx CHUR and yields unrealistic model ages  
B numbers in parenthesis are model ages calculated using typical crustal \(^{147}Sm/^{144}Nd = 0.12 for samples possessing unusually high \(^{147}Sm/^{144}Nd.\)
Fig. 10. Plot of initial $\varepsilon_{\text{Nd}}$ versus $\varepsilon_{\text{Sr}}$ for representative samples of the East Coast Province granitoid associations. The field for the mafic rocks is from chapter five and these rocks are discussed in that chapter in more detail. Small curves link duplicate analyses.
Plutons belonging to the peraluminous S-type association of the East Coast Province have $\varepsilon_{\text{Nd}}$ from -3.5 to -5.6 and $\varepsilon_{\text{Sr}}$ from +60 (0.709) to +155 (0.715). There is a significant overlap of the fields for the S- and I-type associations despite their distinct chemical and mineralogical differences. Highly felsic phases of the S-type plutons however possess abnormally high $^{147}\text{Sm} / ^{144}\text{Nd}$ ratios of 0.16 to 0.19 compared to the range of 0.10 to 0.13 shown by the I-type and mafic S-type plutons. The very high ratio of 0.194 (sample 128) indicates a flat REE pattern which is unusual for rocks of granitoid composition.

Miller and Mittlefehldt described some examples of highly felsic granitoids with low REE abundances, flat or erratic REE patterns and very large negative Eu anomalies similar to these East Coast examples. They suggested that such features may be attributed to monazite/sphene/allanite fractionation or high temperature magmatic-hydrothermal fluid interaction that may still yield relatively 'normal' granitoid textures.

$T^{\text{Nd}}_{\text{DM}}$ and zircon inheritance ages

Zircon inheritance ages for the plutons of the East Coast Province plutons are not as well constrained by comparison with those of the West Coast Province. The two available inheritance patterns indicate an age range in the region 800 to 1300 Ma. Although more results are required to test the consistency of these figures, the inherited zircon ages are consistent with $T^{\text{Nd}}_{\text{DM}}$ ages of 900 to 1400 Ma for the I-type granitoids and 1180 Ma for a mafic S-type granodiorite. Some highly fractionated felsic S-type plutons, although showing normal $\varepsilon_{\text{Nd}}$ values of -4 to -6, are characterized by abnormally high $^{147}\text{Sm} / ^{144}\text{Nd}$ ratios approaching the chondritic figure of 0.1967. Magmatic or hydrothermal processes have evidently fractionated Sm and
Nd in these rocks. Resulting $T_{DM}^{Nd}$ ages are unreasonably high (2-3 Ga).

Using a 'typical' granitoid value of 0.12, the resulting $T_{DM}^{Nd}$ ages are 1000 - 1300 Ma and are more realistic model age estimates.

An interesting observation which needs to be substantiated by more work is that sample 132 (an I-type granodiorite) contains inherited zircons and possesses a 'crustal' initial $\epsilon_{Nd}$ value of -6 whereas sample 120 lacks inherited zircons and has an initial $\epsilon_{Nd}$ value of -1.3 that is suggestive of 'mantle' derivation. Both samples, however, have very similar initial $\frac{\text{Sr}^{87}}{\text{Sr}^{86}}$ ratios of 0.708 and 0.707 respectively. In addition, both display quite similar $\delta^{18}O$ values of +9.5 and +8.1 (H.P. Taylor, pers. comm.). A possible albeit speculative interpretation is that I-type granitoids like sample 132 are derived from old (Precambrian) zircon-bearing igneous protoliths which are LREE-enriched but have low Rb/Sr ratios whereas those like sample 120 are derived from melting of young underplated basaltic crust followed by limited assimilation of old lower crust.

The zircon and $T_{DM}^{Nd}$ age data for both I-type and S-type plutons show a crude overlap in the region 900 - 1300 Ma, which may be interpreted to represent the 'crust formation' age of the East Coast Province crustal segment. This age range is significantly younger than the estimate for the West Coast Province and argues for a younger basement and source regions for the plutonic suites in the East Coast Province. An alternative interpretation is that the age of the basement is better estimated using the S-type granitoids ($T_{DM}^{Nd} = 1100 - 1300$ Ma) and the I-type granitoids with the most negative $\epsilon_{Nd}$ values ($T_{DM}^{Nd} \approx 1100 - 1400$ Ma). This interpretation implies that the 800 Ma upper intersection of the zircon discordia chord for sample 132
is an underestimate of the true age of the old zircons due to pre-Permian Pb loss whereas the upper intersection of the zircon discordia chord for sample 106 gives a good estimate of the age of the source zircons. The 'crust formation' age based on the second interpretation is 1100 - 1400 Ma, a range not significantly different from that based on the first interpretation.

**Other Phanerozoic batholith terrains**

Despite their complexities, there is general agreement between zircon inheritance ages and $T_{DM}^{Nd}$ ages for the Peninsular Malaysian batholiths. Both systems have been used to estimate 'crust formation' ages for the West Coast Province (~1500 - 1700 Ma) and the East Coast Province (~1100 - 1300 Ma). Recent U-Pb zircon and Nd data are available for other continental margin granitoid batholith terrains. These include the Lachlan Fold Belt granitoids of southeastern Australia (Williams, 1977; McCulloch and Chappell, 1982), Caledonian granitoids of the British Isles (Pidgeon and Aftalion, 1978; Hamilton et al., 1980) and western U.S.A. (DePaolo, 1981b). In all cases, I have recalculated the data to both the depleted mantle evolution model used in this study and the Nd normalization value adopted at the A.N.U. laboratory to facilitate comparisons.

For the southeastern Australian S-type granitoids, the complex zircon inheritance patterns give upper intersections in the range 1500 - 1900 Ma. These inheritance patterns indicate the presence of old zircon components dominated by a population having this ~1500 - 1900 Ma age range (Williams, 1977). $T_{DM}^{Nd}$ ages (McCulloch and Chappell, 1982) ranging from 1400 to 1650 Ma are in close agreement with the zircon results. Complex reverse discordia patterns for zircons from the I-type plutons do not yield reliable estimates of the ages of the
old components but $T_{DM}^{Nd}$ ages of $\sim 1600$ Ma are given by I-type plutons with the most negative $\varepsilon_{Nd}$ values. This figure is within the range of $T_{DM}^{Nd}$ ages calculated for the S-types.

Pidgeon and Aftalion (1978) observed that the general trend of zircon discordia patterns for the Caledonian granitoids of the British Isles have upper intersection ages of $\sim 1500$ - $1600$ Ma. Two exceptions were ages of 1320 Ma and 1050 Ma for the Ben Vuirich and Helmsdale masses respectively. $T_{DM}^{Nd}$ ages calculated from the study by Hamilton et al. (1980) give ages of 1500 - 1800 Ma for the S-type 'Older' and 'Newer' Granites and $\sim 1300$ - 1500 Ma for probable I-type 'Newer' Granites with the most negative $\varepsilon_{Nd}$ values (Foyers, Etive Complex, Ben Nevis).

Work by DePaolo (1981b) and DePaolo and Farmer (1982) on the Mesozoic-Tertiary batholith terrain of southwestern United States indicate that two basement provinces were tapped by the magmas. A southeastern province with $\varepsilon_{Nd} \sim -11$ and an older northwestern province with $\varepsilon_{Nd} \sim -18$ were proposed. Exposed metasedimentary country rocks yield comparable values of -11 to -18. $T_{DM}^{Nd}$ estimates of basement ages vary from 1200 - 1800 Ma. Zircons from the granitoids have generally not yielded reliable upper intersections because of their near-concordant nature. However, zircon work by Silver (1971) have revealed ages of $\sim 1700$ - 1800 Ma for the oldest widespread Precambrian units that outcrop in the region.

For the four terrains discussed, we can at the very least conclude that, despite their complexities, regional zircon inheritance ages are comparable to $T_{DM}^{Nd}$ ages. This highlights the utility of the Nd method for identifying basement ages especially for plutons with no inherited zircon. A more significant conclusion, however, is that
where \( T_{DM}^{Nd} \) ages correspond to regional inherited zircon ages, we appear to be obtaining 'crustal formation' ages of juvenile segments of continental crust extracted from the mantle in the past. Granitoids, can in this manner, be used as images of underlying crustal segments and offer a unique way of mapping and identifying otherwise hidden basement blocks and estimating the 'formation' ages of such blocks.

**Nd ISOTOPIC CHARACTERISTICS OF PHANEROZOIC CONTINENTAL MARGIN MAGMATIC BELTS**

On an \( \varepsilon_{Nd} - \varepsilon_{Sr} \) diagram (Fig.11) Phanerozoic continental margin magmatic belts of various ages define a relatively narrow curving band that flattens out in the vicinity of \( \varepsilon_{Nd} \approx -10 \) to \(-12\). S-type granitoids typified by the West Coast Province of Peninsular Malaysia, Koscuisko Batholith of southeastern Australia and the British Caledonian 'Older' Granites display patterns that reflect the old metasedimentary basement from which they were derived. These patterns are characterized by elevated and variable initial \( \varepsilon_{Sr} \) values and a more limited initial \( \varepsilon_{Nd} \) variation. I-type magmatic belts like those of the Andean, southwestern United States and the Bega Batholith of southeastern Australia show a much smaller range in \( \varepsilon_{Sr} \) but large \( \varepsilon_{Nd} \) variations. This is due to the complex interaction between mantle- and crustal-derived magmas that produce I-type batholiths and their associated volcanic belts. Consequently, the most negative initial \( \varepsilon_{Nd} \) values in such batholiths would best approximate the range reflecting the regional basement.

A plot of \( \varepsilon_{Nd} \) against time, shown in Fig.12, displays both the range of initial \( \varepsilon_{Nd} \) variation in the Phanerozoic plutonic-volcanic belts plotted in Fig.11 and the generalized emplacement/extrusive ages for these belts. The three reference Nd
Fig. 11. Initial $\varepsilon_{Nd}$ and $\varepsilon_{Sr}$ values of Phanerozoic continental margin granitoid belts. IA: primitive arc basalts, PR: Peninsular Ranges, SN: Sierra Nevada, AD: Andean, BCN: British Caledonian Newer granites, LI: Lachlan I-types, MEC: Malaysian East Coast, FH: French Hercynian, LS: Lachlan S-types, BCO: British Caledonian Older Granites, MWC: Malaysian West Coast, HM: Himalaya.

Fig. 12. Plot of $\varepsilon_{Nd}$ versus time showing the relationships of the evolution line for CHUR, depleted mantle and Archaean crust. Vertical solid bars denote range in $\varepsilon_{Nd}(T)$ for each plutonic (or volcanic) province. Horizontal locations of solid bars denote crude emplacement (or extrusive) ages. Symbols as in Fig. 10. Stippled band denotes evolution band for the range in $\varepsilon_{Nd}(T)$ values for crustal end-members of the plutonic provinces as indicated by the most negative $\varepsilon_{Nd}(T)$ value of individual provinces.
evolution lines in Fig.12 are (1) that for a 2500 Ma crust with a "typical" continental crust \( ^{147}\text{Sm}/^{144}\text{Nd} \) ratio of 0.12, (2) that representing CHUR evolution (horizontal line) and (3) that representing the depleted mantle evolution model outlined in the introduction. Archaean crust (> 2500 Ma) would be characterized by present-day \( ^{143}\text{Nd}/^{144}\text{Nd} \) exceeding \( \varepsilon_{\text{Nd}} \approx -25 \). The most negative present day \( \varepsilon_{\text{Nd}} \) value found in Phanerozoic continental margin granitoids is \( \approx -16 \). It is clear that none of the voluminous Phanerozoic continent margin batholiths have tapped old Archaean crust to any significant extent. It was argued previously that the agreement of zircon inheritance ages and \( T_{\text{DM}}^{\text{Nd}} \) ages for terrains where both methods have been used to study granitoids suggest that \( T_{\text{DM}}^{\text{Nd}} \) ages are reasonably reliable estimates of 'crust formation' ages. Extending this argument to other batholith terrains, then \( T_{\text{DM}}^{\text{Nd}} \) ages of the crustal end-members of batholiths represented by the intersections of the stippled band in Fig.12 with the depleted mantle evolution line yield 'crust formation' ages between \( \approx 1300 - 1800 \) Ma. The dominance of crustal end members with present day \( \varepsilon_{\text{Nd}}(0) \) values between -9 to -16 must indicate a widespread distribution of continental crust with such isotopic signatures.

**CONCLUSION**

Nd, Sr and zircon U-Pb data for granitoid batholiths in Peninsular Malaysia have been used to estimate 'crustal formation' ages of the sutured crustal blocks present. In both cases, old Proterozoic crust is clearly indicated and the ages of the blocks are estimated as \( \approx 1500 - 1700 \) Ma for the West Coast Province and \( \approx 1100 - 1300 \) Ma for the East Coast Province. An examination of other batholith terrains for which both Nd and zircon U-Pb data are
available also reveal comparable $T_{\text{DM}}^{\text{Nd}}$ ages and regional zircon inheritance ages. Such agreement suggests that zircon-bearing granitoid magmas were directly extracted from the mantle in the Precambrian or that the time interval between mantle extraction, intracrustal melting and addition to the crust as granitoid material was relatively short (≈ 100-200 Ma). In either case, reliable 'crustal formation' ages can be obtained from Nd and zircon studies.

'Crustal formation' ages for crustal segments underlying many important Phanerozoic continent margin magmatic belts were estimated using $T_{\text{DM}}^{\text{Nd}}$ ages and all fall in the general range 1300 - 1800 Ma. The available granitoid data therefore suggest that the mid-Proterozoic was a very important episode of addition of juvenile crustal material from the mantle. Evaluation of the pervasive nature of this episode and its general implications for crustal growth will have to await data for major crustal segments tapped by batholiths in east and central Asia, U.S.S.R., the Middle East, western Canada, and eastern United States.

Based on the present data available, the absence of Archaean Nd signatures at continental or former continental margins is striking. The lower and middle crust tapped by voluminous continental margin plutonic-volcanic belts in many important regions show consistently mid-Proterozoic ages and is highly suggestive that little Archaean crust extends beneath the Proterozoic and Phanerozoic cover of present day continental accretion margins. This predominance of mid-Proterozoic ages at continent margins implies that many of the continental blocks may have reached their present size by the mid-Proterozoic and that continental accretion processes by Phanerozoic arc-type magmatism may have added little to the growth of continents.
There is substantial evidence for the episodic evolution of continental crust and presently, the questions that have arisen are: when were the times of crustal additions and how much new crust was formed during each episode. Patchett (1983), for example, refers to major episodes at 0.7 Ga, 1.8 Ga, 2.8 Ga and >3.0 Ga. Many workers have proposed that a significant portion of the continental crust was stabilized by the end of the Archaean. Combined Nd and zircon inheritance work could allow a more quantitative assessment of this proposal, notably with respect to hidden basement blocks. Granitoid data suggest an important mid-Proterozoic crustal addition episode(s) although its importance relative to the 2.5 - 3.0 Ga episode cannot as yet be assessed.

Hofmann and White (1982) and Ringwood (1982) recently proposed mantle geodynamic models to explain the fate of subducted oceanic lithosphere. These workers suggested that subducted oceanic crust accumulates deep in the mantle, probably near the 650 km discontinuity, for timescales of 1 - 2 Ga. Thermal instabilities developed during this period cause partial melting, buoyant diapiric ascent and convective overturns in the deep mantle source regions. Ringwood (1982) suggests that rapid growth of continental crust may be associated with such episodes of instability. It is tempting to speculate that major crust-formation episodes at 2.5 - 3.0 Ga (McCulloch and Wasserburg, 1978; Taylor, 1979) and ~1.3 - 1.8 Ga as suggested by the results of this study are related to such episodic mantle instabilities.
CHAPTER SEVEN
SUMMARY AND SYNTHESIS

A summary of the main conclusions reached in previous chapters of this thesis and a discussion of the salient features of the Palaeozoic-Mesozoic granitoid plutonism of Peninsular Malaysia in relation to the general problem of granitoid magmatism are presented in this chapter.

Geochronology of the batholiths

Results of the U-Pb zircon study and a reexamination of the Rb-Sr whole rock data of Bignell and Snelling (1977) indicate that the granitoids of the West Coast Province are of late Triassic age. The 198-220 Ma age range obtained from the U-Pb zircon study is consistent with the interpretation that the oldest K-Ar mica ages of Bignell and Snelling (1977) are approaching emplacement ages. A number of Rb-Sr whole rock linear arrays yielding apparent ages older than Triassic were obtained by Bignell and Snelling (1977). The geochemical and isotopic data discussed in chapters three and four of this thesis appear to confirm the presence of plutons with such anomalously old apparent Rb-Sr ages. In the light of the unambiguous late Triassic U-Pb zircon ages for these plutons, it is suggested that these old Rb-Sr whole rock ages are the result of assimilation of country rocks with highly radiogenic Sr during the ascent and emplacement of the granitoid magmas. Assimilation can produce arrays that yield apparent ages older than the true emplacement age because of the slope generated on a $^{87}\text{Sr}/^{86}\text{Sr} - ^{87}\text{Rb}/^{86}\text{Sr}$ correlation diagram even before the whole rock system starts recording age information caused by the in situ decay of $^{87}\text{Rb}$. The variable Sr initial ratios of the granitoids in the West Coast Province may also reflect the influence
of assimilation processes when isotopic differences between magma and assimilated materials are not large. If combined assimilation-fractionation processes are as common as a number of workers (e.g. Taylor, 1980; DePaolo, 1981a; James, 1982) contend, the accuracy and usefulness of the Rb-Sr whole rock isochron method need to be reassessed.

Results of the U-Pb zircon study of plutons from the East Coast Province substantiate the Permian to Triassic ages proposed by Bignell and Snelling (1977) based largely on their K-Ar mica results and a small number of Rb-Sr whole rock determinations.

Inherited Precambrian zircon populations are present in the granitoids of both the West Coast and East Coast Provinces. This feature indicates that both provinces are underlain by or were adjacent to Precambrian continental crust. Upper intersections of zircon discordia patterns for the West Coast Province are in the range 1500-1700 Ma and these are significantly older than the 800 Ma and 1350 Ma upper intersections for two East Coast Province plutons.

Granitoid batholiths of the West Coast Province

The unambiguous late Triassic ages indicated by the U-Pb zircon study and the results of more equivocal Rb-Sr work support the suggestion of Mitchell (1977) that the West Coast Province granitoids can be related to the late Triassic continent collision event which led to the creationization of mainland Southeast Asia.

The West Coast Province granitoids studied are exclusively S-type i.e. derived by partial melting of metasediments. The composite batholiths and complexes in this S-type province are divisible into a number of structurally continuous, mineralogically related and
chemically and isotopically coherent suites. Each suite appears to correspond to an individual upper crustal fractionation series.

Two distinct S-type groups may be distinguished in the West Coast Province - (1) a peraluminous, high-Al biotite + muscovite + Mn-rich garnet group and (2) a metaluminous-moderately peraluminous group with the mafic assemblage low-Al biotite + sphene + actinolitic hornblende. The presence of sphene and the higher Fe$_2$O$_3$ contents in the ilmenites of the second group indicate that they were derived from relatively oxidized magmas. Despite these differences, both groups possess similar S-type chemical characteristics, such as narrow ranges of SiO$_2$ composition, high K$_2$O/Na$_2$O ratios, high to very high abundances of Rb, Th, U, Pb, Y, Sn and Cs, high initial $^{87}$Sr/$^{86}$Sr ratios, constant and negative initial $\varepsilon_{Nd}$ values and high $\delta^{18}$O (H.P. Taylor, pers.comm.).

Major element oxide and Rb-Ba-Sr patterns of the West Coast Province granitoids are best explained by crystal fractionation processes. The chemical trends can be modelled by a process of plagioclase-biotite - quartz (+ zircon, apatite, ilmenite, monazite) fractionation in intermediate-upper crustal magma chambers. Very high Rb (> 500 ppm) and low Ba (< 50 ppm) and Sr (< 50 ppm) abundances in leucogranites are especially diagnostic. Fifty to seventy percent fractionation of parental granodioritic magmas is required to achieve the trace element signatures of many of the leucogranites. More extensive fractionation is indicated if such parental granodioritic magmas had undergone some degree of prior differentiation before emplacement at the present level of exposure. As discussed previously, concurrent assimilation and fractionation processes are indicated by anomalously old Rb-Sr whole rock isochron ages and highly
variable initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The magmas differentiated towards $H_2O$-saturated conditions and the evolution of a discrete fluid phase enriched in rare metals and volatiles. Reaction with such magmatic fluids produced anomalous chemical signatures in a number of the high-$SiO_2$ granites.

Sn enrichment in the West Coast Province granitoids is the result of initial high abundances of Sn in the magmas, the relatively reduced nature of the magmas, their extended fractionation history and most importantly, the enhancement in the fluid phase by volatile complexing.

High initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (between 0.71 to 0.74), negative initial $\epsilon_{\text{Nd}}$ values (-6 to -10) and inherited zircon features indicate derivation of the granitoids from old metasedimentary crust.

**Batholiths of the East Coast Province**

The Permian - Triassic plutonism of the East Coast Province of Peninsular Malaysia can be related to eastward subduction along a trench now represented by the medial Bentong-Raub ophiolite line. Three distinct plutonic associations can be recognized - a dominant I-type granitoid association, a minor S-type granitoid association and a mafic association. The plutonism in the East Coast Province is distinctly bimodal. There is a $SiO_2$ break from 55 to 65% $SiO_2$ that reflects the scarcity of rocks of intermediate compositions.

The I-type granitoids range in composition from diorite to granite but are mainly of granodioritic composition. The mafic mineralogical assemblage is low-Al biotite + hornblende + sphene + magnetite + ilmenite. A coexisting sphene + magnetite assemblage is
indicative of the high $f_{O_2}$ nature of some I-type granitoids but there exists a spectrum from such oxidized plutons to highly reduced plutons possessing near stoichiometric ilmenite without coexisting magnetite.

Mafic silicate compositional variations, major element oxide trends and Rb-Ba-Sr relationships indicate that plagioclase-hornblende-biotite-quartz fractionation (possibly with concurrent assimilation) in high level crustal magma chambers was the dominant magma differentiation process producing the observed diversity in granitoid compositions. The occurrence of I-type leucogranites with peraluminous chemistries and high Ba, Rb and low Sr abundances is especially diagnostic of extended crystal fractionation.

In common with many continent margin granitoid batholiths, the East Coast Province I-type granitoids display isotopic characteristics intermediate between those of typical mantle-derived magmas and old continental crust. Initial $^{87}$Sr/$^{86}$Sr ratios range from 0.705 to 0.710 but are typically around 0.707-0.708, initial $\varepsilon_{Nd}$ values vary between 0 to -6.5 and $\delta^{18}O$ values are typically between +7 and +9 (H.P. Taylor, pers.comm.).

Plutons recognized as belonging to an S-type association range from granodiorite to granite in composition. The ferromagnesian assemblage is high-Al biotite + ilmenite + muscovite. S-type chemical and isotopic characteristics include peraluminous compositions, high $K_2O/Na_2O$, low $Fe_2O_3/FeO$, high Rb, Th, U, Pb, Y, Cs, high initial $^{87}$Sr/$^{86}$Sr ratios (0.709-0.715), negative $\varepsilon_{Nd}$ initials (-3.5 to -6) and high $\delta^{18}O$ (>9). Extensive plagioclase-biotite-quartz fractionation produced much of the observed chemical trends and yielded distinctive highly fractionated leucogranites with
high Rb and very low Ba and Sr contents.

Units comprising the mafic association in the East Coast Province include primitive olivine gabbro complexes, orthopyroxenite-norite complexes, gabbro-diorite complexes and mafic dyke swarms. Mafic units are minor, making up less than five percent of the area of the batholiths. Mineralogical and chemical features of the mafic rocks are similar to those of orogenic basalts and andesites with broad tholeiitic to high-K tholeiitic affinities. High initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.706 to 0.709) and negative initial $\varepsilon_{\text{Nd}}$ values (-2.5 to -4.5) are suggestive of derivation from enriched mantle sources.

The clear tripartite division of plutonic associations is a fundamental feature of the East Coast Province. The S-type and mafic associations represent magmas derived from old metasedimentary crust and from the mantle respectively. It is suggested that the best way to explain the isotopic features of the I-type granitoids, the presence of the 55-65% SiO$_2$ gap and the lack of fractionation relationships between mafic rocks and I-type granitoids is derivation primarily by partial melting of a heterogeneous lower crust consisting of juvenile underplated and old crustal components of mafic-intermediate compositions. Minor hybridization with mantle-derived magmas and hybridization-assimilation of metasedimentary components are likely to be important. The latter would explain the high $\delta^{18}$O values of some I-type plutons. However, the clear chemical and mineralogical separation of I- and S-type granitoids argues against the possibility that S-type magmas are derived from parental I-type magmas that incorporated large amounts of sedimentary components. Intermediate 'I-S' types required by this
mechanism are not observed and the S-type suite indicate extended fractionation histories from primitive peraluminous S-type magmas. The suggestion that S-types grade into I-type 'roots' (e.g. Hildreth, 1981) does not seem compatible with the data for the East Coast Province.

The bimodal separation of mafic rocks and granitoids of the East Coast Province by an intermediate $SiO_2$ gap may be attributed to the relatively rapid migration of plutonic loci. The 10-20 Ma duration for a plutonic locus in the East Coast Province probably did not facilitate extensive hybridization between mantle-derived magmas and crustal melts to allow production of large volumes of intermediate magmas like the ubiquitous tonalites of many continent margin batholiths.

Some broad implications of aspects of the granitoid plutonism in Peninsular Malaysia

Regional tectonics: The juxtaposition of two contrasting batholith provinces documented in this thesis — an older subduction-related Permian-Triassic province and a younger collision-related late Triassic province — is elegantly predicted by plate tectonic theory. Closure of an ocean or marginal basin by subduction of oceanic lithosphere results in the collision of two originally separate continental blocks. The existence of two crustal blocks in Peninsular Malaysia is further substantiated by the isotopic results obtained in the present study. Zircon inheritance results and Nd isotopic data indicate an older (~1500 – 1700 Ma) western crustal block and a younger (~1100 – 1300 Ma) eastern crustal block.
Two contrasting plutonic provinces: A first order difference between the West Coast and East Coast plutonic provinces is the exclusively S-type nature of the former. This contrasts with the expanded compositional nature of the East Coast Province where I-type and S-type granitoids and mafic rocks are all recognized. The scarcity of volcanic rocks in the West Coast Province contrasts with the abundance of volcanic rocks with an expanded compositional range similar to the plutonic rocks in the East Coast Province. Compositional differences of the granitoids of both provinces largely reflect the S-type nature of the West Coast Province granitoids and the predominantly I-type nature of the East Coast Province granitoids. This is notably apparent from the fields of the granitoids on the modified ACF plots of Fig.1. The field for the West Coast Province granitoids lies above the plagioclase-biotite tie-line whereas the bulk of the East Coast Province granitoids plot in the metaluminous region below the plagioclase-biotite tie-line. The high contents of incompatible LIL elements like Rb, Th, U, Sn, Pb, Cs together with other distinctive chemical signatures of the S-type granitoids of the West Coast Province are highlighted in Fig.2. The high incompatible element abundances is a consequence of both derivation from metasediments with high contents of these elements and extended crystal fractionation.

Melt generation processes: The absence of mantle-derived magmas in the West Coast Province and the emplacement of large volumes of S-type magmas adjacent to the suture where maximum crustal thickening is indicated by the results of the gravity traverse by Van Klinken and Ho (1982) suggest that ultrametamorphism of underthrusted metasedimentary sequences during continent collision is the most probable mechanism for generation of the West Coast Province granitoids. For the case of the East Coast Province, the presence of mantle-derived mafic units is
Fig. 1 Plots of East Coast and West Coast granitoids on modified ACF diagrams.
Fig. 2A. Selected Harker variation diagrams showing high abundances of some LIL elements in the West Coast Province (WCP) granitoids compared to the East Coast Province (ECP) granitoids.
Fig. 2B Selected variation diagrams illustrating the contrasting chemical characteristics of the West Coast and East Coast Province granitoids.
Fig. 2C Plots of Rb versus Sr and La vs. Th showing the high Rb/Sr and low La/Th of the S-type granitoids of the West Coast and East Coast Provinces compared to the I-types of the East Coast Province.
suggestive of a mantle heat source despite their minor exposed volumes. A basaltic 'heat trigger' that causes partial melting of mafic-intermediate lower crust and metasedimentary crust at higher levels would explain the association of both I- and S-type granitoids in the East Coast Province. The incapability of normal geotherms to reach the high temperatures required to generate I-type tonalitic-granodioritic liquids and the need for a mantle heat source were discussed by Wyllie et al. (1977). A restite unmixing model (White and Chappell, 1977) that regards tonalitic magmas as mixtures of mafic restite and lower temperature felsic liquid would lessen the need for excessive amounts of mantle-derived magmas to act as a heat source. Amphibole of Mg$_{50}$ compositions on the liquidus of melts of dacitic composition is indicated by petrographic and geochemical studies. This suggests high temperatures of ~900°C (e.g. Gill, 1981) and the extensive fractionation signatures also imply high temperature non-minimum melts.

Although the I-type granitoids display isotopic features intermediate between the mantle-derived mafic units and the S-type granitoids, simple two end-member mixing or assimilation-fractionation cannot explain the detailed isotopic and chemical variations. This is consistent with the views of workers like Taylor (1980) and Hildreth (1981) who emphasized that heterogenous sources, deep-seated hybridization and polybaric fractionation-assimilation in source regions which may have vertical extents of kilometres, are more realistic.
Magma differentiation: The main conclusion from a consideration of the chemical characteristics of the Peninsular Malaysian granitoids is that crystal fractionation in intermediate-upper crustal magma chambers and variable separation of increasingly silicic differentiated liquids best explain the diversity of I- and S-type granitoids. This process was supplemented by concurrent assimilation, limited hybridization and reaction with magmatic-hydrothermal fluids to various degrees. Restite unmixing whereby a melt is progressively separated from its refractory residue does not seem to be important at the present level of emplacement. Assimilation may be a widespread phenomenon but chemical signatures of such a process seem to have been swamped by crystal fractionation signatures. Assimilation is, however, indicated by a number of anomalously old Rb-Sr whole rock isochron apparent ages and variable \(^{87}\text{Sr}/^{86}\text{Sr}\) initial ratios. For the East Coast Province I-type granitoids, intermediate to high \(^{18}\text{O}\) values displayed by many plutons which still retain their distinctive I-type mineralogy and major element chemistries are also suggestive of assimilation effects. Differentiation towards \(\text{H}_2\text{O}\) - saturated conditions and late-stage separation of an aqueous, volatile-enriched fluid phase is most strongly displayed in the West Coast Province.

An important point that needs to be stressed, however, is that the above discussion is largely relevant to upper crustal differentiation processes only. Chemical signatures representing differentiation processes at deep crustal levels are largely lost and complications arising from eruptive losses, if any, cannot presently be deciphered. Hildreth (1981) suggested that many plutons do support volcanic fields and documented the volcanic evidences for complexly zoned magma chambers generated by purely liquid state differentiation processes. He argued that eruptive losses make subvolcanic signatures
uninterpretable and believed that more reliable insights into plutonic systems were allowed by volcanics that sample differentiating magmas at a single instant in time. Thorpe and Francis (1979) and Atherton et al. (1979) had, however, previously cautioned against using volcanic rocks to image underlying plutonic processes. These workers showed that even for the well-exposed Peruvian Andes, volcanic rocks, ring complexes and granitoid batholiths seem to exist as discrete entities - there are no unambiguous links between plutonic and volcanic rocks and geological and geochemical evidence indicate time-composition differences between plutonism and volcanism.

At deep crustal levels and regions of partial melting, restite unmixing as described by White and Chappell (1977) can very likely be an important way to produce variations in granitoid compositions. Products of partial melting (both melt and refractory residue) may move up 'en masse' or with variable retention of residue to higher crustal levels.

Although the chemical variations documented for the Peninsular Malaysian granitoids are best explained by crystal fractionation processes at the present levels of exposure, the evidences are compelling that highly contrasting magmatic behaviour may dominate at different depths. Differing conclusions for different terrains need therefore not be considered contradictory. Caution should be exercised in attempting to extend differentiation processes observed for a single, well documented terrain to other terrains. By the same token, processes documented for volcanic-subvolcanic systems and granitoids associated with deep-seated metamorphic complexes may differ significantly from those controlling granitoid emplacement at intermediate crustal depths. At any one time, only a single level of
exposure can be studied and understood for a granitoid system with a gradional, multi-stage, polybaric history. The work of Wyborn et al. (1981) is particularly interesting in this respect. These workers suggested that plutonic and volcanic suites could be recognized and paired as single volcano-plutonic associations in the Lachlan Fold Belt. In addition, xenoliths found in some of the volcanic suites were interpreted by these workers as representing restite, (i.e. refractory residue from partial melting). Such studies which attempt to decipher volcanic, plutonic and source characteristics related to individual magmatic suites can potentially yield valuable insights into the multi-stage histories of granitoid systems.

Variations in batholith provinces and an I-S spectrum

I-type granitoids have been distinguished in this study as those derived primarily from igneous or metaigneous sources that may or may not be of first stage mantle origin, whereas S-type granitoids are derived primarily from sedimentary sources that have undergone an extended period of weathering and sedimentary recycling. It was further shown in this thesis that considerable chemical and isotopic variations exist within each granitoid type. In this section, salient features of I- and S- type granitoids found in various tectonic environments are reviewed and the significance of the chemical and isotopic variations and gradations between the various environment associations are discussed.

Five first-order granitoid environment associations were suggested in chapter one. The schematic diagram illustrating their idealized tectonic locations is reproduced in figure 3. Although the Tertiary examples of granitoids found in the various environments suggested are rather unambiguous, it is stressed that the pre-Tertiary
### Island Arc

**为例：** New Britain

- **I-type plutons**
- **岩浆岩类型**：辉长岩到花岗岩
- **侵入岩**：无
- **同源火山岩**：无
- **花岗岩体**：无

### Accretionary Wedge

**为例：** southwest Alaska

- **I-, S-type plutons**
- **岩浆岩类型**：辉长岩到花岗岩
- **侵入岩**：无
- **同源火山岩**：无
- **花岗岩体**：无

### Continent Margin Plutonic Arc

**为例：** Coastal Batholith of Peru

- **I-, S-type plutons**
- **岩浆岩类型**：辉长岩到花岗岩
- **侵入岩**：无
- **同源火山岩**：无
- **花岗岩体**：无

### Continent Margin Plutonic Belt

**为例：** Idaho Batholith, Western U.S.A.

- **S-type plutons**
- **岩浆岩类型**：辉长岩到花岗岩
- **侵入岩**：无
- **同源火山岩**：无
- **花岗岩体**：无

### Intracontinental Plutonic Belt

**为例：** Idaho Batholith, Western U.S.A.

- **S-type plutons**
- **岩浆岩类型**：辉长岩到花岗岩
- **侵入岩**：无
- **同源火山岩**：无
- **花岗岩体**：无

### Continent Collision Plutonic Belt

**为例：** Tertiary leucogranites

- **S-type plutons**
- **岩浆岩类型**：辉长岩到花岗岩
- **侵入岩**：无
- **同源火山岩**：无
- **花岗岩体**：无

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**Fig. 3** Highly schematic environment settings for granitoids of the orogenic association. Tectonic relationships are highly idealized. Refer to chapter one for a discussion of complexities and qualifications.
examples may, in some instances, be more equivocal.

1. Granitoid plutons associated with intraoceanic island arc settings like New Britain (Whalen, 1980) and Aleutian Islands (Perfit et al., 1980; McCulloch and Perfit, 1981) can be identified as I-type. These granitoid are typically of diorite-tonalite-mafic granodiorite compositions and are often associated with gabbroic units. Very low abundances of elements such as Rb, Th, U, Pb and Cs are a notable feature, as are the flat or slightly LREE-enriched REE patterns. Isotopic features are very similar to those of arc basalts and these include mantle $\delta^{18}O$ values of $\sim 5 - 7$, low initial $^{87}Sr/^{86}Sr$ ratios of $0.703 - 0.705$ and positive initial $\varepsilon_{Nd}$ values of $\sim 5 - 10$.

2. Granitoid plutons found intrusive into the accretionary wedge complex at subduction settings are present in the southern Alaskan Sanak-Baranof belt (Hudson et al., 1979; Hill et al., 1981), the 'Outer Zone' of southwest Japan (Ishihara, 1978; Shibata and Ishihara, 1979) and, possibly, the New England Fold Belt of eastern Australia (Shaw and Flood, 1981). Granitoids of both I- and S-type, together with associated gabbros, can be recognized in such settings. I-type granitoids in the New England Fold Belt display initial $^{87}Sr/^{86}Sr$ ratios of $0.703$ to $0.706$, initial $\varepsilon_{Nd}$ values of $0$ to $+6$ and $\delta^{18}O$ typically between $+7$ and $+9$ (McCulloch et al., 1982; Shaw and Flood, 1981). A striking feature is the primitive Sr and Nd isotopic characteristics of the associated S-type plutons despite their high $\delta^{18}O$ values ($> +10$). Initial $^{87}Sr/^{86}Sr$ ratios are typically $0.706 - 0.707$ and initial $\varepsilon_{Nd}$ values are around $-1$ to $+3$. 
3. Granitoid batholiths formed along the margins of old continents display the most complex chemical and isotopic patterns. Examples include the Sierra Nevada Batholith of western U.S.A., the Coastal Batholith of Peru and the Bega-Berridale Batholiths of eastern Australia. The East Coast Province batholiths of Peninsular Malaysia can be included in this group. I-type and S-type suites together with gabbros are generally present. Granitoids recognized as I-types are mainly of tonalite–granodiorite composition and possess variable initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from 0.703 to 0.710, variable initial $\varepsilon_{\text{Nd}}$ values from +5 to negative values typical of old continental crust (-5 to -10) and $\delta^{18}O$ values typically around +7 to +9. S-type granitoids display isotopic features that reflect their derivation from old continental crust i.e. initial $^{87}\text{Sr}/^{86}\text{Sr}$ > 0.709, initial $\varepsilon_{\text{Nd}}$ more negative than -5, $\delta^{18}O$ > +9 and zircon inheritance patterns.

4. Intracontinental and continent–collision granitoid suites are either dominated by or consist exclusively of S-type granitoids. Examples of the former are the Inner Cordilleran S-type Belt of western U.S.A. (Miller and Bradfish, 1981) and possibly, the Koetong suite of southeastern Australia (Price, 1982). The Himalayan leucogranites (Dietrich and Gansser, 1981) are Tertiary examples of continent collision–related granitoids. The West Coast Province granitoids of Peninsular Malaysia are also interpreted to belong to this group. Both intracontinental and continent–collision granitoid groups show continental crustal signatures as discussed for S-type granitoids of group (3). Partial melting in the root zones of tectonically thickened continental crust produced either by crustal shortening or
underthrusting of continental crust at collision settings can yield low temperature S-type magmas. Mantle heat sources are not essential at such settings.

The preceding examples serve to illustrate the manner in which features of granitoid plutonism from settings where mantle signatures dominate (island arcs) grade into those where the chemical and isotopic signatures are exclusively those of continent crust. The wide range of gradational chemical and isotopic features displayed especially by continent margin batholiths reflect the complex interactions between continental crust and mantle in such terrains. A spectrum of I-type and S-type granitoids is inevitably generated as a result of varying degrees of crust-mantle interactions. The spectrum may result from source region complexities like heterogeneous sources with vertical extents of kilometres and melting of sources with mixed I-S characteristics like volcaniclastic sediments or reflect purely magmatic processes like wall rock assimilation and hybridization. A single $\varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}}$ diagram (Fig. 4) serves to illustrate the isotopic spectrum of previously recognized I- and S-type granitoids. The widespread recognition of an I-S granitoid division in many plutonic provinces (Chappell and White, 1974; Beckinsale, 1979; Pankhurst, 1979; Takahashi et al., 1980; Dietrich and Cansser, 1981; Shaw and Flood, 1981) primarily reflects the strongly bimodal nature of the I-S spectrum in many individual plutonic provinces. Fig. 5 attempts to portray on a schematic two-dimensional plot, the nature of the I-S spectrum in terms of model sources, variations of selected chemical and isotopic characteristics of these sources and tries to represent how examples of a number of I- and S-type suites identified previously would fit on the schematic plot. It is well realized that
Fig. 4 Initial $\varepsilon_{\text{Nd}}$ vs. $\varepsilon_{\text{Sr}}$ plot showing the wide range of isotopic characteristics observed for both S- and I-type granitoids previously recognized. Data for southeastern Australian granitoids after McCulloch et al. (1982).
Fig. 5 attempts to generalize a large number of variables on a single diagram. Nevertheless, it is felt that the usefulness of the generalizations outweighs the risk of oversimplicity. Fig. 5 provides broad perspectives that may be useful to the understanding of the diversity of granitoids and to lessen some misconceptions about the I-S division of granitoid types, especially with regard to the degree of adherence to the features displayed by the I- and S-type suites first distinguished in southeastern Australia.

Granitoids as images of continental crust

Granitoid batholiths sample large volumes of crust and as such, provide bulk images of middle and lower crust. Sm-Nd and zircon U-Pb studies of granitoids can provide constraints on the nature and evolution of these deep crustal regions. Because of the relative immobility of the REE and the difficulty in fractionating Sm from Nd in the crustal environment, Nd isotopes have proved useful for estimating the times when crustal materials were extracted from the mantle. U-Pb zircon inheritance ages, on the other hand, yield estimates of the ages of an ultimate zircon-bearing granitoid source or high-grade metamorphic basement source. Agreement between Nd model ages and zircon inheritance ages would imply that extraction of juvenile crust from the mantle and formation of zircon-bearing granitoid magma represent a single event or were not separated by an extended time interval (< 100 - 200 Ma). The application of these principles was explored in chapter six.

The results of an integrated U-Pb zircon and Nd isotopic study have allowed first order age estimates of the continental blocks underlying the East Coast and West Coast Provinces of Peninsular Malaysia to be made. The results indicate 'crust formation' ages of ~
Fig. 5 Generalized two-dimensional diagram illustrating the possible interactions between the main sources for granitoids that can produce a spectrum of S- and I-type granitoids. Axes are schematic and are not meant to portray quantitative interactions between the various systems. Examples plotted are discussed in the text and references cited. Degree of I- and S-type 'nature' is qualitatively illustrated by the right hand vertical axis. The recognition of separate S- and I-type associations in many continent margin batholith settings is suggestive that the I-S spectrum is strongly bimodal in nature for such terrains.
1500 - 1700 Ma and \( \sim 100 - 1300 \) Ma for the basement blocks of the West Coast and East Coast provinces respectively. The result for the West Coast Province should prove to be an important constraint in attempts to solve the problem of the identity of the parent Precambrian craton from which western Southeast Asia was detached in the Jurassic.

A significant observation is the consistent mid-Proterozoic 'crust formation' ages of crustal segments underlying continent margin fold belts as indicated by the Nd and U-Pb zircon data for continent margin volcano-plutonic belts. Widespread mid-Proterozoic 'crust formation' ages along continent margins and the absence of Archaean signatures imply that many continental blocks had reached their present sizes by the mid-Proterozoic. Although the existing data base is not extensive, the ubiquity of mid-Proterozoic 'crust formation' ages for many major volcano-plutonic belts is striking and suggestive of an important global mid-Proterozoic crustal addition episode. This interpretation, if correct, supports crustal growth models that invoke episodic massive addition/stabilization of juvenile continental crust. Evidence discussed in chapter six led to the suggestion that two major periods of juvenile crustal addition/stabilization can be recognized, \( \sim 2.5 - 3.0 \) Ga and \( \sim 1.3 - 1.8 \) Ga. The bulk of the continental crust was stabilized by 1.3 Ga. Phanerozoic arc magmatism does not appear to be a major mechanism of continental crustal growth.
APPENDIX 1 : THE U-Pb ZIRCON METHOD

(A) Fundamentals

The general principles on which the U-Pb zircon method is based have been discussed by various workers including Faure (1977) and Gebauer and Grunenfelder (1979). In this brief review, only selected aspects pertinent to the understanding of the data obtained in this thesis are discussed.

The U-Pb zircon method uses the radioactive decay of $^{238}\text{U}$ to $^{206}\text{Pb}$ and $^{235}\text{U}$ to $^{207}\text{Pb}$ for age determinations. Measured isotopic compositions of Pb in zircon are related to U by the familiar decay equations:

$$
\left( \frac{206\text{Pb}}{204\text{Pb}} \right)_{\text{meas}} = \left( \frac{206\text{Pb}}{204\text{Pb}} \right)_{\text{initial}} + \left( \frac{238\text{U}}{204\text{Pb}} \right) (e^{\lambda_1 t} - 1) \tag{1}
$$

$$
\left( \frac{207\text{Pb}}{204\text{Pb}} \right)_{\text{meas}} = \left( \frac{207\text{Pb}}{204\text{Pb}} \right)_{\text{initial}} + \left( \frac{235\text{U}}{204\text{Pb}} \right) (e^{\lambda_2 t} - 1) \tag{2}
$$

where $t$ = age since time of crystallization of the zircon for a closed system involving no loss or gain of either U or Pb,

$\lambda_1, \lambda_2$ = decay constants of $^{238}\text{U}$ and $^{235}\text{U}$ respectively.

In addition, a $^{207}\text{Pb}/^{206}\text{Pb}$ age can be calculated by combining equations (1) and (2) to yield

$$
\frac{^{207}\text{Pb}}{^{206}\text{Pb}} = \frac{235\text{U}}{238\text{U}} \left( \frac{e^{\lambda_2 t} - 1}{e^{\lambda_1 t} - 1} \right) \tag{3}
$$

where $235\text{U}/238\text{U}$ is a constant equal to 1/137.88.

For closed U-Pb zircon systems (i.e. concordant systems), all three ages should agree within analytical error. In most cases however, the three ages calculated do not coincide and the apparent
Fig. 1.1 Concordia diagram illustrating the nature of the discordance patterns for (1) the simple case of variable radiogenic Pb loss from different zircon fractions of a cogenetic population, and (2) the simple case of incorporation of old zircon components in younger magmatic zircon populations. Ages marked by ticks on concordia curve are in Ma.
ages often increase in the order $t_{206} < t_{207} < t_{207-206}$. This discordant behaviour indicates that the assumption of closed system behaviour is not fulfilled. The interpretation of discordant U-Pb zircon ages was largely solved by the use of the concordia diagram (Wetherill, 1956). The concordia curve (Fig.1.1) on such a diagram represents the locus of all present day $\frac{206\text{Pb}}{238\text{U}}$ and $\frac{207\text{Pb}}{235\text{U}}$ ratios for closed U-Pb systems, where

$$ \frac{206\text{Pb}}{238\text{U}} = \left[ \frac{(206\text{Pb})_{\text{meas}} - (206\text{Pb})_{\text{initial}}}{(204\text{Pb})_{\text{meas}}} \right] / \left[ \frac{238\text{U}}{204\text{Pb}} \right] $$

A similar expression holds for the $\frac{207\text{Pb}}{235\text{U}}$ system.

The nature of the processes causing U-Pb zircon systems to be discordant can be recognized and the effects of these processes sometimes overcome by plotting a number of cogenetic zircon fractions on a concordia diagram. Complex discordant patterns observed in U-Pb zircon systems can be explained in terms of two main processes - loss of radiogenic Pb and zircon inheritance (i.e. survival of old zircons in a young granitoid sample).

A number of mechanisms are believed to be responsible for Pb loss in zircon systems. These include post-emplacement metamorphic resetting, low temperature alteration, breakdown of the zircon structure due to time-integrated radiation damage and continuous diffusion of Pb from zircon crystals. Pb loss causes points to plot below the concordia curve. Such points have U-Pb ages younger than the true age of zircon crystallization. The discordance line defined by cogenetic zircons that have suffered different amounts of Pb loss may, however, be extrapolated to the point of zero Pb loss (i.e. the upper intersection of the discordance line with the concordia curve) to obtain the zircon crystallization age (case1, Fig.1.1).
The normal discordance pattern discussed above contrasts with the pattern resulting from the effects of zircon inheritance. Zircon inheritance yields points that also plot below the concordia curve but such points have individual U-Pb ages older than the true age of zircon crystallization. Assuming that the pattern is not complicated by superimposed Pb loss, the discordance line produced by mixing between a magmatic component and a single inherited zircon component yields a lower intersection with the concordia curve that corresponds to the crystallization age of the magmatic zircon component (Case 2, Fig.1.1).

More complicated scenarios involving multi-episodic Pb loss, combined Pb loss and zircon inheritance features and incorporation of complex populations of old zircons are discussed by Gebauer and Grünendolder (1979).

(B) Chemical procedures

Zircon concentrates were obtained after crushing to <500 \( \mu \text{m} \) by conventional Wifley table, heavy liquid and magnetic separation techniques. These concentrates were then separated into size fractions and handpicked to obtain pure zircon splits based on morphological and colour differences. U-Pb zircon chemistry were based on the methods described by Krogh (1973). Handpicked zircons were washed in hot concentrated 7N - HNO\(_3\) before dissolution. High pressure dissolution of zircons in Teflon capsules using HF-HNO\(_3\) were carried out in bombs held at \( \sim 200^\circ \text{C} \) for durations of 1 - 2 weeks. Solutions were converted to chloride, spiked using a \( ^{235}\text{U} \) tracer and divided into two portions. The smaller portion was spiked with a \( ^{208}\text{Pb} \) spike tracer whereas the larger portion was retained for the determination of Pb isotopic composition. U and Pb were
separated using AG1-X8 anion exchange resin by HCl elution. All acids were prepared by sub-boiling distillation (Mattinson, 1971) in teflon stills. Laminar flow heads provide dust-free environments for clean handling of reagents and samples. The total Pb blank was typically 0.3 ng per analysis.

(c) Mass spectrometry

Pb was loaded for mass spectrometry on single Re filaments using the H₃PO₄ - silica gel technique. U was loaded in H₃PO₄ onto a Ta₂O₅ film on a single Re filament and measured as UO⁺. Measurements were made on a modified Nuclide Analysis Associates 30.5 cm radius, 60° sector mass spectrometer linked on-line to a HP – 1000 computer. Isotopic ratios were measured by ABAB - type rapid peak switching using the magnetic field.

Pb loads were raised slowly to reach ion beams of 10⁻¹¹ amps before data collection was started. A data set consists of 25 one-second integrations of alternate ion beams. 24 ratios of two isotopes obtained in this manner constitute a set and 20 such sets constitute a Pb run. ²⁰⁴Pb/²⁰⁶Pb ratios were measured using the Faraday cup by turning up to the maximum achievable signal at the final stages of a Pb run.

The 'static zero' measurement was made at mass 202.5 for Pb and a 'dynamic zero' correction is added to the 'static zero' to correct for the finite response time of the electrometer after measuring a large signal. The dynamic zero is estimated as the difference between measurements at the background position during static and dynamic modes.
The Pb program allows for an examination of the extent of fractionation of light relative to heavy isotopes occurring during a Pb run. A plot of $^{208}\text{Pb}/^{206}\text{Pb}$ versus time is displayed and the choice of the time when all ratios are normalized to is decided by the operator depending on the correlation observed. For most runs however, there were no significant differences between $^{208}\text{Pb}/^{206}\text{Pb}$ ratios measured at the start and those at the end of a run.

(D) Data evaluation and calculation of U-Pb ages

All Pb isotopic data listed in this thesis have been corrected for a 0.3 ng total Pb blank and common Pb. The isotopic composition of the laboratory reagent blank is $^{206}\text{Pb}/^{204}\text{Pb} = 17.97$; $^{207}\text{Pb}/^{204}\text{Pb} = 15.55$ and $^{208}\text{Pb}/^{204}\text{Pb} = 37.71$. Common Pb corrections used in age calculations are $^{206}\text{Pb}/^{204}\text{Pb} = 18.8$; $^{207}\text{Pb}/^{204}\text{Pb} = 15.8$ and $^{208}\text{Pb}/^{204}\text{Pb} = 39.1$; these being based on the analyses of Malaysian galenas reported by Bignell and Snelling (1977). Contributions to any error in ages arising from uncertainties in common Pb corrections are minimal for the data reported in this study. For the case of the lowest measured $^{206}\text{Pb}/^{204}\text{Pb}$ ratio (365 for sample 120, +170 concordant fraction - table 1, chapter three), if the data were corrected using 1500 Ma common Pb ($^{206}\text{Pb}/^{204}\text{Pb} = 16.1$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.4$, $^{208}\text{Pb}/^{204}\text{Pb} = 35.8$) calculated using model III of Cumming and Richards (1975), the $^{206}\text{Pb}/^{238}\text{U}$ ages increases from 227.2 Ma to 228.9 Ma and the $^{207}\text{Pb}/^{235}\text{U}$ age increases from 227.4 Ma to 229.5 Ma. For this worst case result, the age difference is of the same order as the estimated analytical error of $\pm$ 2-3 Ma. For the general case of measured $^{206}\text{Pb}/^{204}\text{Pb}$ ratios $> 600$, ages calculated for any concordant zircon fraction using the galena common Pb values and 1500 Ma common Pb values differ by less than 1 Ma. Ages obtained by intersections of discordance lines with
the concordia curve possess errors that are much greater than errors
due to uncertainties in common Pb corrections.

The results presented in this thesis have been calculated using
the decay constants and isotopic abundances recommended by Steiger and
Jäger (1977). Where applicable, discordance lines have been fitted
using a York (1969) regression program. A correlation coefficient of
0.96 for errors in x and y have been calculated using equation (13) of
Ludwig (1980). Intersection ages were determined by solving for the
intercepts of the best-fit line with the concordia curve.
Uncertainties in intersection ages were obtained by determining the
intercepts when the best-fit line is varied about its centroid by its
2σ uncertainty limits.

Spikes were calibrated against shelf standards of dissolved U
wire and NBS 982 equal atom Pb standard. Uncertainties in spike
calibration are less then 0.2% for both U and Pb. Isotopic ratios
obtained for the radiogenic Pb standard NBS - 983 during the period of
study are listed in table 1 and displayed graphically in Fig.1.2.
208\(^{\text{Pb}}/206\(^{\text{Pb}}\) ratios are identical to NBS values within analytical
error. The Pb isotopic data presented in this study have not been
systematically normalized. 207\(^{\text{Pb}}/206\(^{\text{Pb}}\) and 204\(^{\text{Pb}}/206\(^{\text{Pb}}\) ratios
are slightly lower and higher respectively compared to NBS quoted
values but are comparable within analytical error to the range of
values obtained by other recent studies (Fig. 1.2).

Because of the limited amounts of the internal zircon standard
available, only one analysis of this standard was carried out. The
results of this analysis (aliquot 6) are compared with the previous
determinations summarized by Williams et al. (1982) in Table 2. The
new results are not significantly different from the means of the
Table 1 Results of determinations of radiogenic Pb standard NBS-983

<table>
<thead>
<tr>
<th></th>
<th>run 1</th>
<th>run 2</th>
<th>run 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>NBS value</td>
<td>0.013617 ± 6 0.013627 ± 6 0.013621 ± 8</td>
<td>0.071138 ± 18 0.071123 ± 18 0.071108 ± 24</td>
<td>0.000394 ± 8 0.000382 ± 6 0.000388 ± 8</td>
</tr>
<tr>
<td></td>
<td>0.013619 ± 24 0.013617 ± 6 0.013621 ± 8</td>
<td>0.071201 ± 40 0.071138 ± 18 0.071108 ± 24</td>
<td>0.000371 ± 20 0.000382 ± 6 0.000388 ± 8</td>
</tr>
</tbody>
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Table 2 U-Pb data for the Myalla Road Syenite zircon standard

<table>
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<tr>
<th>Aliquot</th>
<th>Pb (ppm)</th>
<th>U (ppm)</th>
<th>206Pb ppm</th>
<th>204Pb ppm</th>
<th>238U ppm</th>
<th>235U ppm</th>
<th>206Pb</th>
<th>207Pb</th>
<th>Apparent ages 206Pb 207Pb</th>
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<tbody>
<tr>
<td>5</td>
<td>12.6</td>
<td>397.6</td>
<td>3092</td>
<td>0.02723</td>
<td>0.1846</td>
<td>0.04918</td>
<td>173.2</td>
<td>172.0</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>12.6</td>
<td>391.0</td>
<td>3122</td>
<td>0.02756</td>
<td>0.1905</td>
<td>0.05012</td>
<td>175.3</td>
<td>177.0</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>12.4</td>
<td>387.7</td>
<td>3139</td>
<td>0.02731</td>
<td>0.1873</td>
<td>0.04974</td>
<td>173.7</td>
<td>174.3</td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>12.3</td>
<td>394.6</td>
<td>3942</td>
<td>0.02674</td>
<td>0.1844</td>
<td>0.05002</td>
<td>170.1</td>
<td>171.8</td>
<td></td>
</tr>
<tr>
<td>20</td>
<td>12.6</td>
<td>394.0</td>
<td>2929</td>
<td>0.02732</td>
<td>0.1864</td>
<td>0.04948</td>
<td>173.8</td>
<td>173.5</td>
<td></td>
</tr>
<tr>
<td>mean (+2σ)</td>
<td>12.7</td>
<td>402.8</td>
<td>2333</td>
<td>0.02723 ± 60</td>
<td>0.1866 ± 50</td>
<td>0.04971 ± 78</td>
<td>173.2 ± 3.4</td>
<td>173.7 ± 3.8</td>
<td></td>
</tr>
</tbody>
</table>

Analyses 5, 12, 13, 18, 20 after Williams et al. (1982)
Fig. 1.2 Comparison of determinations of NBS 983 with quoted values and other available determinations.
previous determinations at the 2σ uncertainty level. $^{207}\text{Pb}/^{206}\text{Pb}$ agree to within 1.1% and $^{206}\text{Pb}/^{238}\text{U}$ agree to within 1.9% of the mean values of Williams et al. (1982).

Uncertainties expected from typical internal precision of mass spectrometer runs and uncertainties in spike calibrations alone are 0.2% for Pb-Pb and 0.4% for U/Pb ratios. The data for the internal standard suggest higher uncertainties but these may reflect inhomogeneity of the aliquots analysed and thus represent maximum error estimates. Actual uncertainties lie between the theoretical minimum and these maximum error estimates. Uncertainties adopted in this study are 0.3% and 1.0% for $^{207}\text{Pb}/^{206}\text{Pb}$ and U/Pb respectively at the 2σ level.
APPENDIX 2 : THE Sm-Nd METHOD

(A) Fundamentals

In this thesis, the Sm-Nd method has been used both for geochronological purposes and as a petrogenetic tracer, but in a way quite different from that for the well known Rb-Sr system. The usefulness of the Sm-Nd system and the systematics of Nd isotopic variations in crustal and mantle rocks have been discussed by DePaolo and Wasserburg (1979) and McCulloch and Wasserburg (1978).

Sm and Nd are members of the REE group. $^{147}\text{Sm}$ decays to $^{143}\text{Nd}$ by $\alpha$-decay. This process has a half-life of $1.06 \times 10^{11}$ years. Nd has seven isotopes including $^{143}\text{Nd}$, these are $^{142}\text{Nd}$, $^{144}\text{Nd}$, $^{145}\text{Nd}$, $^{146}\text{Nd}$, $^{148}\text{Nd}$ and $^{150}\text{Nd}$. By convention, the increase in $^{143}\text{Nd}$ caused by decay of $^{147}\text{Sm}$ is normalized to $^{144}\text{Nd}$. $^{144}\text{Nd}$ is radioactive ($\alpha$-decay to $^{140}\text{Ce}$) but has a long half-life of $2.1 \times 10^{15}$ years and may be regarded as essentially stable for present purposes.

$^{143}\text{Nd}/^{144}\text{Nd}$ as measured today is related to $^{147}\text{Sm}$ by the familiar decay equation

$$
\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_{\text{meas}} = I + \left(\frac{^{147}\text{Sm}}{^{144}\text{Nd}}\right)(e^{\lambda t} - 1) \tag{1}
$$

where $t$ is the time of closure of the system to movement of Sm and Nd,

$\lambda$ is the decay constant of $^{147}\text{Sm} (= 6.54 \times 10^{-12} \text{ yr}^{-1})$.

$I$ is the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of the system.

Variations in $I$ allow a characterization of the sources of crustal rocks to be made. The strength of the Sm-Nd technique lies, however, in the contrasting reaction of the Sm-Nd system to crustal
processes compared to the Rb-Sr system. Whereas Rb and Sr and strongly fractionated by crustal melting, magmatic differentiation, metamorphism and sedimentary recycling, Sm and Nd are not significantly affected by such processes because of the relative immobility of the REE and the coherent behaviour of Nd and Sm. This resistance to resetting by crustal processes has proved useful for the determination of the times when Nd and Sm were fractionated during a mantle melting event and chemically fractionated mantle-derived materials were added to the continental crust. The basis for this approach is explained below.

Nd systematics of terrestrial rocks are best discussed by reference to three model reservoir systems - (1) a CHUR (chondritic uniform reservoir) mantle with flat REE patterns and an unfractionated $^{147}\text{Sm}/^{144}\text{Nd}$ value of 0.1967, (2) a continent crustal reservoir with LREE - enriched patterns and $^{147}\text{Sm}/^{144}\text{Nd} < $ chondritic, typically $\sim 0.11 - 0.13$, and (3) a depleted mantle (DM) reservoir as represented by MORB sources with $^{147}\text{Sm}/^{144}\text{Nd} > $ chondritic. The depleted mantle reservoir is widely regarded to represent the depleted residue produced after extraction of continent crustal components from a CHUR - type mantle (e.g. Allègre, 1982).

The $^{143}\text{Nd}/^{144}\text{Nd}$ of CHUR - type mantle through time is given by the expression:

$$(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}(t) = (^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}(0) + (^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}}(e^{\lambda t} - 1)$$

where $$(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}(0) = 0.511836$$ and $$(^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} = 0.1967$$

Because of the usefulness of a 'bulk earth' chondritic reference, the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of a rock at time T is widely expressed as deviations from CHUR rather than expressed as the numerical value
itself. The $\varepsilon_{\text{Nd}}(T)$, or deviation from CHUR in parts in $10^4$, of a rock is given by:

$$\varepsilon_{\text{Nd}}(T) = \left[ \frac{^{143}\text{Nd}/^{144}\text{Nd}}{^{143}\text{Nd}/^{144}\text{Nd}}_{\text{CHUR}} - 1 \right] \cdot 10^4 \quad (3)$$

Fig. 2.1 illustrates how $\varepsilon_{\text{Nd}}$ varies with time for (1) a CHUR mantle with $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$, (2) a depleted mantle with a model $^{147}\text{Sm}/^{144}\text{Nd} = 0.225$, and (3) a LREE-enriched reservoir representing continental crust with $^{147}\text{Sm}/^{144}\text{Nd} = 0.12$.

**Model ages**: Fig. 2.1 will form the basis of the following discussion on Nd model ages. Two types of model ages can be considered for crustal rocks originally derived from the mantle. Both require different assumptions, and depending on the evolution and fractionation history of the mantle source, can give different model ages.

The first type of model age is termed the $T_{\text{Nd CHUR}}$ age and its calculation requires the assumption that a crustal sample was originally derived from a source with $^{147}\text{Sm}/^{144}\text{Nd}$ equal to that for CHUR mantle. For this case, the $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}}(T) = (^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}(T)$ for time $T$ when fractionation of Sm and Nd and inferred mantle extraction occurred. Using this relationship, $T$ can be obtained by rearranging equations (1) and (2) to give:

$$T_{\text{Nd CHUR}} = \frac{1}{\lambda} \ln \left[ 1 + \frac{^{143}\text{Nd}/^{144}\text{Nd}}{^{147}\text{Sm}/^{144}\text{Nd}}_{\text{meas}} - \left( ^{143}\text{Nd}/^{144}\text{Nd} \right)_{\text{CHUR}}^{0} \right] \quad (4)$$

In Fig. 2.1, the intersection of the Nd evolution line of a sample calculated using equation (1) and the horizontal CHUR evolution line represents the $T_{\text{Nd CHUR}}$ age.
Fig. 2.1 Plot of $\varepsilon_{Nd}$ versus age showing the horizontal CHUR evolution line and representative evolution lines for depleted mantle and continental crust. The plot also illustrates graphically how $T_{\text{CHUR}}$ and $T_{\text{DM}}^{Nd}$ ages are obtained.
The second type of model age is termed the $\frac{Nd}{DM}$ age. Its calculation assumes derivation of a crustal sample from a depleted mantle with $^{147}$Sm/$^{144}$Nd > 0.1967. For the hypothetical case of a depleted mantle derived originally from CHUR mantle of 2700 Ma and having a present day $^{143}$Nd/$^{144}$Nd represented by MORB, $T$ is obtained from the equation

$$T_{\text{DM}}^{\text{Nd}} = \frac{1}{\lambda} \ln \left[ 1 + \frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{meas}} - (^{143}\text{Nd}/^{144}\text{Nd})_{\text{DM}}^o}{(^{147}\text{Sm}/^{144}\text{Nd})_{\text{meas}} - (^{147}\text{Sm}/^{144}\text{Nd})_{\text{DM}}^o} \right]$$

where $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{DM}}^o = 0.512348$ and $(^{147}\text{Sm}/^{144}\text{Nd})_{\text{DM}} = 0.225$

In Fig.2.1, the intersection of the Nd evolution line, calculated using equation (1), of a sample and the model depleted mantle evolution line represents the $T_{\text{DM}}^{\text{Nd}}$ age.

There is still some uncertainty as to the nature and extent of the involvement of CHUR and DM mantle reservoirs in the generation of continental crust. Model ages calculated using equations (4) and (5), however, allow first order estimates of crustal formation ages to be made. In general, $T_{\text{CHUR}}^{\text{Nd}}$ ages may be regarded as minimum ages of the times of crustal formation as there is increasing evidence that continental crust was extracted from previously depleted mantle during the Phanerozoic (DePaolo, 1981c; McCulloch and Chappell, 1982; this thesis).

$\varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}}$ correlations: Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of rocks can also be expressed in the $\varepsilon$-notation analogous to that for the Nd system. The expression is:

$$\varepsilon_{\text{Sr}}(T) = \left[ \frac{(^{87}\text{Sr}/^{86}\text{Sr})_{\text{sample}}}{(^{87}\text{Sr}/^{86}\text{Sr})_{\text{CHUR}}} - 1 \right] \cdot 10^4$$

where $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{CHUR}}(T)$ is obtained from

$$(^{87}\text{Sr}/^{86}\text{Sr})_{\text{CHUR}}^o = (^{87}\text{Sr}/^{86}\text{Sr})_{\text{CHUR}}^T + (^{87}\text{Rb}/^{86}\text{Sr})_{\text{CHUR}}(e^{\lambda T} - 1)$$

and $(^{87}\text{Sr}/^{86}\text{Sr})_{\text{CHUR}}^o = 0.7048$; $(^{87}\text{Rb}/^{86}\text{Sr})_{\text{CHUR}} = 0.085$
$^{87}\text{Sr}$ calculated in this fashion was used for discussions of the suggested correlation between the $\text{Sr}$ and $\text{Nd}$ isotopic systems. The early results indicating that $\varepsilon_{\text{Nd}}$ and $\varepsilon_{\text{Sr}}$ values of modern day basalts from mid-oceanic ridges, ocean islands and primitive arcs were correlated and defined a 'mantle array' were discussed by DePaolo and Wasserburg (1979). As continent crustal contamination could be discounted for many samples studied, the 'mantle array' was interpreted to have resulted from the mixing of CHUR mantle and depleted mantle. The $\varepsilon_{\text{Nd}} - \varepsilon_{\text{Sr}}$ diagram, as summarized in Fig.2.2, represents a useful tool to discuss the origins of terrestrial igneous rocks in terms of derivation from depleted mantle, CHUR mantle, juvenile crust and old crust.

(B) Chemical Procedures

Chemical procedures adopted are those of McCulloch and Chappell (1982). Samples typically weighing $\sim 70 \text{ mg}$ were spiked with $^{150}\text{Nd}$ and $^{147}\text{Sm}$ enriched tracers prepared by Wasserburg et al. (1981). Dissolution was done in Teflon bombs using HF - HC10$_4$. Solutions were converted to the chloride form for ion exchange procedures. Rb, Sr and the REE are separated from each other using Dowex 50 - 8X, 100 - 200 mesh cation exchange resin and HCl as the eluant. Separation of Sm and Nd was achieved using a second column - a capillary tube filled with a 30 cm column of Dowex 50 - 8X, $\geq 400$ mesh resin adjusted to the NH$^+$ form. The REE cut was eluted with 0.2M 2 - methyllactic acid, adjusted to pH = 4.6 so as to optimize separation of Sm and Nd. To increase the rate of passage, the eluant was passed through the column under pressure. New resin was used for every sample. Total processing blanks for Nd are less than 1 ng and are negligible for the samples studied - a typical sample processed contains 2000 - 4000 ng
Fig. 2.2 Schematic plot of present day $\varepsilon_{\text{Sr}}$ versus $\varepsilon_{\text{Nd}}$ showing the positions of the mantle array, primitive arc rocks and old crust as typified by the Lewisian rocks of Scotland. Dashed lines represent generalized $T_{\text{DM}}$ crust formation ages for corresponding $\varepsilon_{\text{Nd}}(0)$ values. Diagram after DePaolo (1981b).
(C) Mass spectrometry

Nd was ran by loading ~1000 ng of Nd on the side filaments of a triple Re filament bead using HCl and analyzed as Nd+. A $^{144}\text{Nd}^+$ ion beam of $10^{-11}$ amps was easily achieved with the technique used. Possible interferences caused by $^{142}\text{Ce}$, $^{144}\text{Sm}$ and $^{150}\text{Sm}$ were checked at masses 140 and 152 respectively before actual data collection commenced. In all cases, interferences were insignificant.

Measurements were made on a 22.85 cm radius, 60° sector mass spectrometer, the MSZ, linked on-line to a HP-1000 computer. Isotopic ratios were measured by ABAB-type rapid peak switching of the magnetic field. A data set consists of 25 one-second alternating integrations of ion beams of two isotopes. About 25 such sets constitute a Nd run. Measurements of $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{150}\text{Nd}/^{144}\text{Nd}$ were bracketed by measurements of $^{146}\text{Nd}/^{144}\text{Nd}$ and $^{142}\text{Nd}/^{144}\text{Nd}$ in such a way that $^{146}\text{Nd}/^{142}\text{Nd}$ obtained from the latter two ratios could be used to monitor mass fractionation occurring during a Nd run. Variations in $^{146}\text{Nd}/^{142}\text{Nd}$ was used for correction of mass fractionation using

the expression:

\[
(\frac{^{143}\text{Nd}}{^{144}\text{Nd}})_{\text{corrected}} = (\frac{^{143}\text{Nd}}{^{144}\text{Nd}})_{\text{meas}} D^{144}
\]

where \(D^4 = \frac{(^{146}\text{Nd}/^{142}\text{Nd})_{\text{meas}}}{0.636151}\)

Measurements of background were made of mass number 140.5. Mass scans were performed to ensure that $^{140}\text{Ce}$ and $^{141}\text{Pr}$ did not effect the background position. $^{143}\text{Nd}/^{144}\text{Nd}$ obtained for BCR-1 over the period of study average 0.511843 ± 20.
Sm was loaded in an identical fashion to Nd and analysed as Sm\(^+\). Concentrations of Sm were obtained using standard isotope dilution equations.

Nd isotopic as well as Nd concentration and Sm concentration determinations were made on a single totally spiked sample. \(^{143}\text{Nd}/^{144}\text{Nd}\) can be measured to better than 5 parts in \(10^5\) and the uncertainty in \(^{147}\text{Sm}/^{144}\text{Nd}\) is estimated to be better than 0.5%.

Alternatively, because equation (1) is that of a straight line of the form \(y = mx + c\), a plot of present day \(^{87}\text{Sr}/^{86}\text{Sr\text{\(}} - c\text{\text{\(}}\) against present day \(^{87}\text{Rb}/^{86}\text{Sr\text{\(}} - c\text{\text{\(}}\) for a number of igneous samples that have remained closed systems, yields a straight line

\[ \text{slope} = \frac{\text{initial}}{\lambda} \]

where \(\lambda\) is the age of closure of the system to movement of Sr and Rb.

The statistical treatment of Sr-Sr isochrons followed in this thesis is based on that discussed by Harley et al. (1969).
APPENDIX 3 : THE Rb-Sr METHOD

(A) Fundamentals

The method is based on the β-decay of $^{87}$Rb to $^{87}$Sr which has a half-life of $4.88 \times 10^{10}$ years. An extended account of the fundamentals and applications of the Rb-Sr method may be found in Faure (1977).

Present day $^{87}$Sr/$^{86}$Sr measured in a rock or mineral is related to $^{87}$Rb by the decay equation:

$$\frac{^{87}\text{Sr}}{^{86}\text{Sr}}_{\text{meas}} = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{\text{initial}} + \left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}}\right)_{\text{meas}}(e^{\lambda t} - 1) \quad (1)$$

where $t$ is the age of closure of the system to movement of Rb and Sr, and $\lambda = 1.42 \times 10^{-11} \text{ yr}^{-1}$.

If the age ($T$) of a granitoid sample has been determined by another radiometric method (e.g. U-Pb zircon), the initial $^{87}$Sr/$^{86}$Sr ratio can be obtained by using equation (1) if the present day $^{87}$Sr/$^{86}$Sr and $^{87}$Rb/$^{86}$Sr are known.

Alternatively, because equation (1) is that of a straight line of the form $y = mx + c$, a plot of present day $^{87}$Sr/$^{86}$Sr ($y$ - axis) against present day $^{87}$Rb/$^{86}$Sr ($x$ - axis) for a number of cogenetic samples that have remained closed systems yields a straight line called an isochron whose

slope = $e^{\lambda t} - 1$, and

intercept($c$) = $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_{\text{initial}}$

The statistical treatment of Rb-Sr isochrons followed in this thesis is based on that discussed by McIntyre et al. (1966). The McIntyre isochron fitting program used in this study yields four model
isochron ages for cases when analysed points cannot be fitted to an isochron within analytical error (i.e. mean square of weighted deviates, MSWD > 1):

(1) a model 1 isochron representing a best-fit line for the data points, the 1σ error and the associated MSWD value.

(2) a model 2 isochron fitted to obtain MSWD = 1 by assuming that the geological and experimental variance is proportional to $^{87}\text{Rb} / ^{86}\text{Sr}$ i.e. $^{87}\text{Sr} / ^{86}\text{Sr}$ variations increase with increasing $^{87}\text{Rb} / ^{86}\text{Sr}$. The model approximates the situation when samples that are analysed have slightly different ages but have identical initial ratios.

(3) a model 3 age fitted with MSWD = 1 by assuming variations in initial $^{87}\text{Sr} / ^{86}\text{Sr}$ independent of $^{87}\text{Rb} / ^{86}\text{Sr}$.

(4) a model 4 age for the situation intermediate between those of Models 3 and 4.

The judgement of which model isochron age is most appropriate in any specific case should be guided by geological considerations together with the isotopic data. In most cases, the various model ages obtained for the granitoids studied are not significantly different.

(B) Experimental techniques

Samples were spiked with $^{84}\text{Sr}$ enriched tracer '746895' and dissolved in Teflon bombs with HF - HC104. Rb and Sr were separated and purified by two passes through a column of Dowex 50 - 8X cation exchange column with HCl as the eluent.

Sr was loaded on single Ta filaments using HCl - $\text{H}_3\text{PO}_4$. At the high current settings for the Sr runs, Rb interference at mass 87 is insignificant. Measured $^{87}\text{Sr} / ^{86}\text{Sr}$ ratios were normalized to
$^{86}\text{Sr}/^{84}\text{Sr} = 0.1194$. Background measurements were made at mass 81.5. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios obtained for NBS 987 on the MSZ mass spectrometer during the period of study average $0.71029 \pm 3$.

The bulk of the samples used in the calculation of isochron ages are those of Bignell and Snelling (1977) who estimated their uncertainties as $+2\%$ for $^{87}\text{Rb}/^{86}\text{Sr}$ and $\pm 0.1\%$ for $^{87}\text{Sr}/^{86}\text{Sr}$.

The dating work of Bignell and Snelling (1977) was limited by the inability to distinguish individual intrusive suites. A reassessment of their results in the light of the recognized division of intrusive suites in the areas studied in this thesis was therefore thought appropriate. The preferred age assignments based on the Rb–Sr whole rock data for individual suites are listed in Table 3.1.
### Table 3.1 Preferred Rb-Sr whole rock age assignments

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Genting Sempah volcanics

### EXPERIMENTAL VARIATION (MODEL 1) ###

| A | 72.66 | B | 662.91 |

**MEAN SQUARE OF WEIGHTED DEVIATES** = 2.19

**F VALUE (50 DEGREES OF FREEDOM FOR X & Y)** = 2.40

**F VALUE (20 DEGREES OF FREEDOM FOR X & Y)** = 2.61

| INITIAL RATIO | 72.664 +/- | 0.015 | AGE | 284.0 +/- | 13.1 M.Y.
|---------------|-------------|-------|-----|-----------|-----|

**EXPERIMENTAL SAMPLE**

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**Bubu Suite (Bintang Batholith)**

### EXPERIMENTAL AND GEOLICAL VARIATION PROPORTIONAL TO AND INDEPENDENT OF X (MODEL 4) ###

| A | 72.66 | B | 0.0012 |

**MEAN SQUARE OF WEIGHTED DEVIATES** = 1.00

**SLOPE OF ABS(DIFF/SE) ON X** = 0.0000249

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**Penanti Suite (Kulin Complex)**

### EXPERIMENTAL VARIATION (MODEL 1) ###

| A | 72.66 | B | 0.0012 |

**MEAN SQUARE OF WEIGHTED DEVIATES** = 71

**F VALUE (50 DEGREES OF FREEDOM FOR X & Y)** = 2.40

**F VALUE (20 DEGREES OF FREEDOM FOR X & Y)** = 2.73

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**EXPERIMENTAL SAMPLE**

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<td>74417</td>
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<td>7 6200</td>
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<td>0 6200</td>
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Kuantan Granite Stock

**** EXPERIMENTAL VARIATION (MODEL 1) ****

\[ A = 71300 \pm 755 \quad B = 80350 \pm 440 \quad \text{R}^2 = 0.625726 \]

Mean square of weighted deviates = 5.21

F value (50 degrees of freedom for X at Y) = 2.78

F value (20 degrees of freedom for X at Y) = 3.11

SE of A = 0.00621

SE of B = 0.000378

INITIAL RATIO = 71309 +/- 0.0197

AGE = 252.3 +/- 5.2 M.Y.

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Direct measurements of X and Y were made in duplicate on UNEU multicoated filter holders. These measurements were based on the calculation of wavelengths of 1000 and 1150 and the corresponding absorption coefficients of the respective samples. The absorption coefficients were compared on the same set of filters with the standard absorption coefficients and the concentration differences are expressed in parts per million (ppm).

Trace elements were determined on both rock and background positions have been determined using the technique described by Norris and Chappell (1977). Rock samples for trace element analysis were matched using pure silica rock as a reference material.
APPENDIX 4: XRF spectrometry

With the exception of FeO (titration), Na$_2$O (flame photometry), H$_2$O$^+$, H$_2$O$^-$ and CO$_2$ (gravimetric) and REE, Cs, Hf, Ta (INAA), all other major and trace elements were determined by X-ray fluorescence spectrometry.

Major elements were measured on fused glass discs and trace elements on pressed powder pellets. Unknown samples were analysed against synthetic standards made from specpure grade chemicals and well calibrated natural rock standards. The XRF data were obtained using Phillips 1220 and 1450 spectrometers.

Direct measurements of Rb and Sr mass absorption coefficients were made in duplicate on small powder discs mounted in Perspex holders. These mass absorption coefficients were used for the calculation of concentrations of trace elements with fluorescent wavelengths between 0.7 - 1.0Å (Rb, Sr, Th, U, Pb, Zr, Nb, Y). Mass absorption coefficients for other trace elements with wavelengths > 1.0Å were calculated from major element analyses and published mass absorption coefficients of the major elements. Relationships between net count rate, mass absorption coefficient and trace element concentration are discussed by Norrish and Chappell (1977, p.242 - 249).

Trace element spectral interferences at both peak and background positions have been corrected using the methods described by Norrish and Chappell (1977). Background correction factors were established using pure SiO$_2$ blanks as described by Chappell et al. (1969).
Details of instrumental conditions and sample preparation techniques adopted for routine major and trace element analyses by XRF are found in Chappell et al. (1969) and Norrish and Chappell (1977).
APPENDIX 5: INSTRUMENTAL NEUTRON ACTIVATION ANALYSIS

Basic principles of the INAA technique are found in the compilation edited by Muecke (1980). The REE, Hf, Ta and Cs data presented in this thesis were determined by INAA using a germanium low-energy photon spectrometer for Ce, Nd, Sm, Eu, Gd, Tb, Yb, Lu, Ta and a lithium-drifted germanium detector for La, Yb, Lu, Ta, Eu, Hf, Cs. All determinations were made against an internal granite standard G-1 that has been calibrated against BCR-1. Details of procedures for sample preparation, irradiation and counting are being prepared in a manuscript by B.W. Chappell. Only a brief summary is offered here.

Powdered samples (0.2 - 0.3g) were weighed into polythene capsules which were then welded. Fourteen unknowns and the internal standard make up an irradiation batch. Known Fe contents of samples were used as neutron flux monitors.

Samples were irradiated at the HIFAR reactor of the Australian Atomic Energy Commission at Lucas Heights, New South Wales, in a thermal flux of \( \sim 5 \times 10^{-12} \text{ n cm}^{-2} \text{ s}^{-1} \) for 24 hours. Samples were counted at the ANU using the LEPS and Ge (Li) detectors after a 1-week 'cooling' period. Data were processed using Ortec amplifiers and accumulated with a 4000-channel Ortec multichannel analyser. Six counts spread over a period of six months were required to obtain the photopeak data for the elements analysed. Isotopes with long half-lives like Cs, Tb, Eu and Gd were counted last so as to minimize interferences from short-lived isotopes. Where more than one photopeak was measured, the average of the determinations was obtained by weighing according to the errors associated with each determination.
Photopeak interference corrections were made using a comprehensive peak stripping program written by B.W. Chappell that corrects for interferences in γ-ray spectra. Precision and accuracy of the INAA technique for the reported elements are estimated to be better than 10%.
APPENDIX 6: Table of major and trace element analyses of the Peninsular Malaysian granitoids obtained by x-ray fluorescence spectrometry.
MALAYSIAN GRANITOIDS

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### Malaysian Granoids

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<td>15.6</td>
<td>19.8</td>
<td>17.6</td>
<td>16.8</td>
<td>14.6</td>
<td>17.6</td>
<td>15.6</td>
</tr>
</tbody>
</table>
APPENDIX 7 : SAMPLE LOCALITIES

Note on granitoid names in Appendix 7

Ideally, granitoids are classified using Streckeisen's (1973) diagram based on modal abundances of quartz : alkali feldspar : plagioclase obtained by point counting of stained rock slabs. This process is however very time consuming. As only a reasonable approximation is required for present purposes, granitoid names have been given on the basis of

(1) thin section evaluation of quartz : alkali feldspar : plagioclase abundances, and

(2) whole rock SiO₂ and CaO contents based on the table below:

<table>
<thead>
<tr>
<th></th>
<th>Diorite</th>
<th>Tonalite</th>
<th>Granodiorite</th>
<th>Adamellite</th>
<th>Granite</th>
</tr>
</thead>
<tbody>
<tr>
<td>% SiO₂</td>
<td>&lt;60</td>
<td>60 - 65</td>
<td>65 - 70</td>
<td>70 - 75</td>
<td>&gt;75</td>
</tr>
<tr>
<td>% CaO</td>
<td>&gt;3</td>
<td>&gt;3</td>
<td>2 - 3</td>
<td>1 - 2</td>
<td>&lt; 1</td>
</tr>
</tbody>
</table>

Sample locations for granitoids from the East coast Province are shown in Fig. 7.1 and 7.2 on the following pages. Locations of samples from the West Coast Province may be found on pages 70a and 72a.
Fig. 7.1 Map showing sample locations in the northern portion of the East Coast Province. Batholith nomenclature is that of Rajah et al. (1977).
Fig. 7.2 Map showing sample locations in the southern portion of the East Coast Province.
GRANITIOIDS OF THE WEST COAST PROVINCE

PENANG ISLAND

1,2  Porphyritic biotite microgranodiorite; Porphyritic muscovite - biotite adamellite
Batu Maung Quarry, Batu Maung, S.E. Penang island. 5° 16' 50"N, 100° 17' 15"E

3,4  Porphyritic biotite leucogranite; prophyritic biotite microgranite
Kheng Kim Quarry, Telok Bahang, N.W. Penang island. 5° 27' 25"N, 103° 13' 5"E

5  Porphyritic biotite adamellite
Chee Seng Quarry, Chee Seng Estate, Tanjung Bungah, N. Penang island. 5° 28' 0"N, 100° 16' 30"E

6  Porphyritic biotite adamellite
Teik Granite Quarry, Vale of Tempe; Tanjung Bungah, Penang 5° 27' 20"N, 10° 16' 40"E

7  Porphyritic tourmaline-biotite granite
JKR Waterfall Quarry, Botanical Gardens, central Penang island. 5° 26' 15"N, 100° 17' 20"E

8  Porphyritic biotite adamellite
Abandoned quarry, Jesselton Heights, central Penang island. 5° 25' 40"N, 100° 17' 50"E
9 Porphyritic muscovite-biotite granite
Soon Gim quarry, base of Penang Hill, Ayer Itam, central Penang island. 5° 24' 40"N, 100° 16' 30"E

10,11 Porphyritic muscovite biotite granite; tourmaline-muscovite microgranite
Paya Terubong Quarry, Paya Terubong, central Penang island. 5° 22' 15"N, 100° 16' 30"E

12 Porphyritic muscovite-biotite microgranite
Kheng Kim Quarry, Paya Terubong - Relau road, Relau, central Penang island. 5° 20' 15"N, 100° 16' 0"E

13 Muscovite-biotite microgranite
Mars Granite Quarry, Kg. Sg. Ara, S.E. Penang island. 5° 19' 0"N, 100° 16' 0"E

14 Biotite adamellite
Choong Lim Lim Quarry, Customs Village, Glugor east Penang island. 5° 22' 10"N, 100° 18' 0"E

Province Wellesley and S. Kedah

15,16,17 Porphyritic biotite adamellite samples. Contain sphene and allanite
Hamir Singh Quarry, Ayer Merah. 9th ml Br.
Mertajam-Kulim road, S. Kedah 5° 21' 0"N, 103° 33' 0"E

18,19 Both samples are porphyritic biotite adamellite containing sphene
Weng Lee Granite Quarry, Berapit, Province Wellesley 5° 21' 30"N, 100° 28' 15"E
Porphyritic biotite adamellite

JKR Quarry, Kg Penanti, Province Wellesley 5° 24' 40"N, 100° 28' 20"E

Muscovite-biotite microgranite

Road cut, 1st ml Bt. Mertajam - Kulim road, Province Wellesley 5° 22' 0"N, 103° 30' 0"E

Biotite granodiorite

Abandoned quarry, 1ml south of Lubok Buntar, S. Kedah 5° 8' 0"N, 100° 34' 55"E

Biotite microadamellite

Lean Seng Chan Quarry, Juru, Province Wellesley. 5° 19' 15" N, 100° 24' 15"E

PERAK

Porphyritic sphene-amphibole-biotite granodiorite ; porphyritic biotite adamellite.

Taiping JKR Quarry, Jln. Asam Kumbang, Taiping, central Perak. 4° 52' 15"N, 100° 44' 0"E

As for 24

Boulders, 1ml southeast of Kg. Changkat Jering, main north-south trunk road, central Perak. 4° 46' 45"N, 100° 44' 0"E

Porphyritic sphene-amphibole-biotite granodiorite.

Boulder along stream at Bt. Berapit, main north-south trunk road, central Perak. 4° 46' 45"N, 100° 49' 0"E
Porphyritic biotite granite (sheared).
Bruas Quarry, 30½ ml from Ipoh along Ipoh-Bruas road,
Bruas, Dindings, central Perak. 4° 31' 10"N, 100° 45'
45"E

SELANGOR AND WEST PAHANG

Porphyritic muscovite-biotite adamellite
Bt. Lanjan Quarry, Damansara, Kuala Lumpur, E.
Selangor. 3° 10' 45"N, 101° 37' 10"E

Tourmaline-muscovite biotite leucogranite
Pernas Construction Quarry, 7th ml Ampang Intake road,
Ampang, S.E. Selangor. 3° 9' 0"N, 101° 47' 0"E

Porphyritic biotite-muscovite granite; muscovite
leucogranite
Private quarry, Taman Cheras Housing Estate, Kuala
Lumpur, S.E. Selangor 3° 6' 20"N, 101° 45' 15"E

Porphyritic tourmaline-muscovite biotite
leucogranite
Outcrop, Taman Cheras Housing Estate, Kuala Lumpur, S.E.
Selangor 3° 5' 40"N, 101° 44' 52"E

Tourmaline-muscovite granite
Construction site, 7th mile Ampang - National Zoo road,
Kuala Lumpur, S.E. Selangor. 3° 11' 30"N, 101° 46' 0"E

Porphyritic biotite granodiorite
Road cut with Italian-Thai inscription, 3½ ml east of
intersection of new highway and old road, new KL - Karak
Highway, E. Selangor. 3° 20' 45"N, 101° 44' 50"E
Porphyritic muscovite-biotite adamellite
Road cut, 5 ml east of intersection of new highway and old road, new KL-Karak Highway, E. Selangor. 3° 21' 20"N, 101° 45' 15"E

Sheared biotite adamellite, contains allanite
Road cut, 9th ml Genting Highlands road, W. Pahang. 3° 25' 0"N, 101° 47' 20"E

Muscovite-biotite granite
Bentong JKR Quarry, Bentong, W. Pahang. 3° 30' 45"N, 101° 52' 0"E

Porphyritic biotite adamellite
Road cut, 3 ml southeast of Rawang, main North-South trunk Road, central Selangor. 3° 17' 45"N, 101° 36' 15"E

Tourmaline-muscovite-biotite granite
Gemilang Quarry, 6ml north-east of Rawang, central Selangor. 3° 23' 40"N, 101° 38' 15"E

Sheared porphyritic biotite adamellite. Contains sphene and allanite.
Wong Koon Quarry, 42 ml Gap Road, N.E. Selangor. 3° 35' 20"N, 101° 42' 10"E

Sheared biotite adamellite
Road cut, 55½ ml Gap Road, N.E. Selangor. 3° 40' 15"N, 101° 44' 50"E

Tourmaline-muscovite microgranite
Road cut, 58½ ml Gap Road, W. Pahang. 3° 41' 50"N, 101° 45' 10"E
44 Porphyritic biotite adamellite
Road cut, 61½ ml Gap Road, W. Pahang. 3° 43' 10"N, 101° 46' 10"E

45 Porphyritic biotite adamellite
Boulder along Sg. Tranum, 64½ ml Gap Road, W. Pahang. 3° 42' 50"N, 101° 46' 30"E

46 Porphyritic biotite adamellite
Road cut, 63½ ml Gap Road, W. Pahang. 3° 43' 10"N, 101° 46' 40"E

47 Porphyritic biotite adamellite
Road cut, 57th ml Gap Road, E. Selangor. 3° 41' 10"N, 101° 45' 0"E

48,49 Tourmaline-biotite microgranodiorite; sheared biotite adamellite
Road cut, ½ ml west of Lentang, new KL-Karak Highway, W. Pahang. 3° 23' 20"N, 101° 53' 0"E

50 Muscovite-biotite adamellite
Road cut, ½ ml east of Lentang, new KL-Karak Highway, W. Pahang. 3° 24' 10"N, 101° 53' 20"E

51,52 Muscovite-biotite adamellite
Garnet-tourmaline-muscovite microgranite
Boulders at ¼ ml south of intersection of new highway and old road, new KL-Karak Highway, E. Selangor. 3° 18' 45"N, 101° 43' 30"E
Porphyritic tourmaline-muscovite biotite adamellite
Quarry at Kg. Batang Benar, 3rd ml Nilai - Pajam road, N. Negri Sembilan. 2° 51' 0"N, 101° 49' 45"E

All porphyritic tourmaline-muscovite-biotite microgranite/microadamellite
Sem Chiu Quarry, 2½ ml Seremban-Mantin road, Seremban, N. Negri Sembilan. 2° 44' 45"N, 101° 55' 0"E

Porphyritic biotite granite
Road cut, 3½ ml Seremban-Mantin road, N. Negri Sembilan. 2° 45' 30"N, 101° 55' 20"E

Sheared tourmaline-muscovite-biotite granite
Sapta Timor Quarry, 8th ml Seremban-Mantin road, N. Negri Sembilan. 2° 48' 0"N, 101° 55' 10"E

Garnet-muscovite-chloritised biotite leucogranite
Boulders, 3rd mile, new Seremban-KL Highway, Seremban, N. Negri Sembilan. 2° 43' 20"N, 101° 54' 40"E

Porphyritic muscovite-biotite adamellite
Eng Lian Quarry, 9th ml Seremban-Tampin road, central Negri Sembilan. 2° 39' 40"N, 102° 1' 40"E

Porphyritic biotite adamellite
Road cut, 2ml from Tampin along Seremban-Tampin road, south Negri Sembilan. 2° 29' 15"N, 102° 12' 0"E
Porphyritic biotite granodiorite
JKR Quarry, 29th mile Seremban-Tampin road, Tampin, south Negri Sembilan. 2° 28' 50"N, 102° 12' 40"E

Porphyritic biotite adamellite
JKR Quarry, Gemenchih, 13th mile Tampin - Gemas road, S. Negri Sembilan. 2° 30' 40"N, 102° 22' 15"E

Porphyritic biotite granodiorite; muscovite-biotite microadamellite
Bt. Mor JKR Quarry; Bt. Mor, Parit Jawa, Muar Selatan, N. Johore. 1° 58' 30"N, 102° 40' 30"E

Porphyritic biotite adamellite. Allanite present.
Goh Ah Soon Quarry, Bt. Panchor, Alor Gajah, N. Malacca. 2° 22' 0"N, 102° 16' 30"E

Both are porphyritic biotite adamellites
Abandoned quarry, Pegoh Estate, Simpang Empat, Alor Gajah, N. Malacca. 2° 26' 30"N, 102° 12' 15"E

Muscovite - biotite granite
Senaling Quarry, Kg. Kuala Pilah, Kuala Pilah, central Negri Sembilan. 2° 41' 55"N, 102° 14' 55"E

Sheared microgranite, sheared biotite granodiorite
Private quarry, Tanjung Ipoh, Kuala Pilah, central Negri Sembilan. 2° 44' 5"N, 102° 12' 50"E

Biotite adamellite
18th Seremban-Kuala Kelawang road, Jelebu, central Negri Sembilan. 2° 53' 0"N, 102° 1' 30"E
Porphyritic rhyolite
Road cut, ½ ml Genting Highlands road, E. Selangor. 3° 21' 20"N, 101° 47' 20"E

Porphyritic rhyolite
Road cut, 1st ml Genting Highlands road, W. Pahang. 3° 21' 30"N, 101° 47' 10"E

Rhyolite
Boulders, 1 ½ ml Genting Highlands road, W. Pahang. 3° 21' 35"N, 101° 46' 50"E

Rhyolite
Road cut, 2 ½ ml Genting Highlands road, W. Pahang. 3° 22' 10"N, 101° 46' 45"E

Pyroxene-biotite microgranodiorite
Road cut, 1 ml west of Kg. Bt. Tinggi, new KL-Karak Highway, W. Pahang. 3° 21' 15"N, 101° 48' 15"E'

As for 78
Road cut, ½ ml west of Kg. Bt. Tinggi, new KL-Karak Highway, W. Pahang. 3° 21' 15"N, 101° 48' 30"E

As for 78
Road cut, 1 ml east of Genting Tunnel, new KL-Karak Highway, W. Pahang. 3° 21' 30"N, 101° 47' 45"E
LATE CRETACEOUS ANOROGenic PLUTONS (except 83)

81  Pink feldspar biotite adamellite
Sagil Quarry, Sagil, Muar Utara, N. Johore. 2° 21' 40"N, 102° 39' 40"E

82  Pink feldspar biotite adamellite
Tangkak Quarry, Tangkak-Bekoh road, Tangkak, N. Johore. 2°19'10"N, 102°31'45"E

83  Quartzo-feldspathic schist
Outcrop along Sg. Mering, north slope of G. Ledang, N. Johore. 2° 25' 0"N, 102° 36' 45"E

84  Pink feldspar biotite adamellite
Bt. Senggeh Quarry, Kg. Selandar, Jasin, E. Malacca. 2° 24' 16"N, 102° 24' 40"E

85, 86, 87  85, 86 are pink feldspar biotite adamellites, 87 is an aplite
Eng Hock Quarry, 21st ml Johore Bahru - Pontian road, S. Johore. 1° 32' 20"N, 103° 32' 50"E

GRANITOIDs AND MAFIC ROCKS OF THE EAST COAST PROVINCE

SINGAPORE & S. JOHORE

88  Porphyritic biotite granodiorite
HDB Mandai Quarry, 20th ml Woodlands Road, Singapore. 1° 24' 0"N, 103° 46' 0"E

89  Gabbro
Associated Granite Quarry, Jln. Perang, central Singapore. 1° 22' 40"N, 103° 45' 30"E
Biotite adamellite; mafic dyke
HDB quarry, Pulau Ubin, island off east Singapore. 1° 25' 0"N, 103° 57' 20"E

Biotite-hornblende granophyre
Aik Hwa Quarry No.2, Pulau Ubin, island off east Singapore. 1° 24' 50"N, 103° 56' 50"E

Biotite-hornblende quartz diorite; biotite-hornblende microadamellite
Gim Huat Quarry, Pulau Ubin, island off east Singapore. 1° 24' 30"N, 103° 57' 40"E

Porphyritic sphene-biotite-hornblende granodiorite
Pulau Sekudu, islet off east Singapore. 1° 24' 15"N, 103° 59' 30"E

Olivine eucrite; eucrite
Linden Hill, Linden Hill Estate, S. Johore. 1° 34' 30"N, 103° 38' 30"E

Biotite granite
Hindhede Quarry, off Woodlands Road, Bt. Timah, central Singapore. 1° 21' 20"N, 103° 46' 30"E

Hornblende orthopyroxenite
Yun Onn Quarry, off Jln. Perang, Bt. Timah, central Singapore. 1° 22' 10"N, 103° 45' 30"E

Biotite granite; porphyritic biotite microgranite
PWD Granite Quarry, off Woodlands Road, Bt. Timah, central Singapore. 1° 21' 40"N, 103° 32' 15"E
Sheared hornblende-biotite granodiorite
Bt. Batok Quarry, off Hillview Road, central Singapore.
1° 21' 10"N, 103° 46' 0"E

Porphyritic biotite-hornblende granodiorite
Kulai Besar Quarry, 19th ml Johore Bahru - Kulai Road,
Kulai, S. Johore. 1° 39' 0"N, 103° 36' 0"E

Hornblende-biotite granodiorite
Teon Choon Quarry, 31st ml Johore Bahru - Ayer Hitam Road, Bt. Batu, S. Johore. 1° 43' 30"N, 103° 27' 40"E

PAHANG, TRENGGANU & KELANTAN

Sphene-biotite-hornblende granodiorite
Boulders, 30th ml Segamat-Kuantan Highway, S. Pahang
Approx. coord. 2° 52' N, 102° 56'E

Porphyritic biotite granite; pegmatitic biotite granite
JKR Quarry, Bt. Ubi, Kuantan, E. Pahang. 3° 49' 0"N, 103° 18' 50"E

Muscovite-biotite granite
Chan Foong Koon Quarry, Bt. Ubi, Kuantan, E. Pahang.
3° 49' 25"N, 103° 18' 40"E

Porphyritic biotite granite
Boulder ½ ml west of 110. 3° 49' 25"N, 103° 18' 30"E

Porphyritic biotite granite (altered)
Boulders at Sg. Lembing, E. Pahang. 3° 54' 25"N, 103° 2' 15"E
Hornblende granite; hornblende granite
Batu Timor Quarry, Tg. Gelang, site of Kuantan Port, E. Pahang. 3° 37' 30"N, 103° 26' 0"E

Biotite granodiorite
Boulders, intersection of old Kg. Air Putih road and new Jerangau – Jabor Highway, Kg. Air Putih, S. Trengganu. 4° 16' 30"N, 103° 13' 15"E

Biotite granodiorite
Road cut, 2km north of Kg. Air Putih, new Jerangau – Jabor Highway. 4° 18' 0"N, 103° 13' 10"E

Biotite-hornblende granite
Kemaman Quarry, 9th mile Kemaman – K. Trengganu road, Kg. Telok Kalong, Kemaman, S. Trengganu.

Hornblende-biotite granodiorite
Abandoned quarry, 11th mile Kemaman – K. Trengganu road, Kemaman, S. Trengganu. 4° 19' 0"N, 103° 29' 15"E

Hornblende-biotite adamellite. Contains allanite.
Abandoned quarry, 1ml north of Kerteh, central Trengganu. 4° 32' 0"N, 103° 27' 45"E

Biotite adamellite
Airport Quarry, Dungun, central Trengganu. 4° 42' 30"N, 103° 26' 10"E

Altered porphyritic biotite microadamellite
Br. Panau JKR Quarry, Tanah Merah, N. Kelantan. 5° 53' 30"N, 102° 10' 30"E
Foliated sheared porphyritic muscovite-biotite granodiorite (West Coast Province)
Boulders, Kg. Batu Gajah, Lanas, W. Kelantan. 5° 48' 40"N, 101° 59' 0"E

Biotite microgranite; biotite adamellite
JKR Quarry, Bt. Buloh, Pulai Chondong, N. Kelantan. 5° 37' 0"N, 102° 15' 30"E

Biotite-hornblende granodiorite
Abandoned Quarry, Bt. Marak, Kg. Peringat, Kota Bharu, N. Kelantan. 6° 2' 20"N, 102° 19' 30"E

Biotite-hornblende granodiorite
Boulders in rubber estate, 18th ml Kota Bharu-Pasir Putih road, Selising, Pasir Putih, N. Kelantan. 5° 54' 30"N, 102° 20' 0"E

Biotite adamellite
Utama Quarry, Kelantan - Trengganu border, Bt. Yong, N. Kelantan. 5° 44' 25"N, 102° 26' 20"E

Muscovite-biotite granite
Bt. Yong Quarry, Bt. Yong, N. Trengganu. 5° 44' 25"N, 102° 26' 20"N

Biotite-hornblende adamellite
Private quarry, Kg. Penarik, Besut, N. Trengganu. 5° 35' 0"N, 102° 47' 30"E

Biotite-hornblende granodiorite
Road cut, Kg. Bt. Bari Besar, south of Kg. Penarik, Besut, N. Trengganu. 5° 33' 40"N, 102° 52' 5"E
Porphyritic biotite adamellite
Kuala Nerus Quarry, K. Trengganu, N. Trengganu. 5° 20' 45"N, 103° 1' 45"E

Porphyritic biotite adamellite
Bt. Jong JKR Quarry, K. Trengganu, N. Trengganu. 5° 21' 0"N, 103° 1' 30"E

Porphyritic biotite adamellite
Hamparan Bumi Quarry, Bt. Cempaka, Bt. Rakit, K. Trengganu, N. Trengganu. 5° 23' 30"N, 103° 1' 45"E

Biotite granodiorite
Tegoh Quarry, Kg. Bt. Sawa, 12th mile K. Trengganu - K. Brang road, central Trengganu. 5° 11' 20"N, 103° 6' 25"E

Altered biotite granite; biotite-hornblende granodiorite
Kenyir Dam Site Construction Quarry, Kg. Jenagor, Ulu Trengganu, central Trengganu. 5° 1' 30"N, 102° 54' 20"E

Biotite granite
Sykt. Trengganu Quarry, 15ml north of Jerangau, Jln. Ajil, Ulu Trengganu, central Trengganu. 5° 1' 30"N, 103° 7' 30"E

Biotite adamellite
Trengganu Tengah Quarry, 32 km north of Kg. Air Putih, new Jerangau - Jabor Highway, central Trengganu. Approx. coord. 4° 36'N, 103° 10'E
Hornblende biotite tonalite

Boulders in Ong Kian Teck Sawmill, 38th ml Kuantan - Temerloh road, E. Pahang. 3° 39' 40"N, 102° 54' 20"E


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