ERRATA

Page 3, line 21. For 'Garner et al., 1976' read 'Garner et al., 1975'.

Page 4, line 5. For 'with an unique' read 'with a unique'.

Page 10, line 3. For 'into one area' read 'into an area'.

Page 13, line 27. For 'as stratigraphic' read 'a stratigraphic'.

Page 21, line 28. For 'axes fellow [ed]' read 'axes follow [ed]'.

Page 23, line 14. For 'usually high' read 'unusually high'.

Page 27, line 19. For 'range in, $^{18}O$' read 'range in $^{13}O$'.

Page 61, line 3. For 'that data' read 'those data'.

Page 68, line 2. For 'with an uniform' read 'with a uniform'.

Page 86, line 24. For 'Examples and' read 'Examples are'.

Page 99, line 7. For 'distribution on the' read 'distribution in the'.

Page 113, line 19. For 'the xenoliths is' read 'the xenolith is'.

Page 124, line 10. For 'grain amounts' read 'grain amounts'.

Page 142, line 27. For 'lower intercepts is' read 'lower intercept is'.

Page 173, line 21. For 'rubidium-strontium' read 'rubidium-strontium'.

Page 175, line 5. For 'old component in' read 'old components in'.

Page 188, line 4. For 'Hewlett packard' read 'Hewlett Packard'.

Page 242, line 14. For 'S_i' read 'S_i/(i-1)'.

Page 245, line 9. For 'S_{ij}' read 'S_{ij}/p(q-1)'.

Page 247, line 13. For 'S_{ij}' read '$S_{ij}/(N-pq)$'.

Page 275, line 2. For '1976:' read '1975:'.

Garner et al., 1975 with an unique ages in excess into one area as stratigraphic axes follow ed usually high range in, $^{18}O$ per mil on I-types that data with an uniform Examples are distribution in the the xenoliths is grain amounts lower intercept is the age of rubidium-strontium old component in Hewlett Packard S_i/(i-1) S_{ij}/p(q-1) S_{ij}/(N-pq) S_{ij}/(N-pq) 1975:
THE BERRIDALE BATHOLITH:

A LEAD AND STRONTIUM ISOTOPIC STUDY

OF ITS

AGE AND ORIGIN

by

Ian Stuart Williams

A thesis submitted in partial fulfilment
of the requirements for the degree of

DOCTOR OF PHILOSOPHY

at the

AUSTRALIAN NATIONAL UNIVERSITY

SEPTEMBER, 1977.
Statement

This thesis is based on experimental work carried out at the Research School of Earth Sciences, Australian National University between January 1974 and September 1977.

Those total-rock rubidium-strontium data which are discussed without acknowledgement are unpublished data of W. Compston and T. Shirahase. All the potassium-argon ages discussed, unless otherwise acknowledged, are unpublished determinations by N.W. Tetley. The assessments of these data and the conclusions drawn from them are the author's own.

With these exceptions, unless it is stated otherwise, the analyses reported in this thesis and the conclusions drawn from them are the author's own.

No part of this work has been submitted to any other university or similar institution.

I.S. Williams.
Acknowledgements

It is my pleasure to acknowledge the considerable help that I have received from others in the course of the present study. In particular, I wish to express my gratitude to my supervisor, Dr. W. Compston, for his guidance of the project and his readiness to assist with the many problems which were encountered as the work progressed. This study has benefited greatly from the high standards which he sets himself and others.

I am deeply appreciative of the close personal interest taken in the present study by Dr. B.W. Chappell; the assistance which he provided through many discussions and in a variety of practical ways was invaluable.

My knowledge of many aspects of the work reported in this thesis has been enhanced by discussions with a number of people in addition to the above: Profs. A.J.R. White, L.T. Silver and P.A.P. Moran, Drs. R.T. Pidgeon, V.M. Oversby, B.L. Gulson, J.J. Foster, I. McDougall, J.N. Boland and R.W. Page and Mrs. L. Wybom.

Messrs M. Cowan and R. Rudowski provided instruction on the techniques of sample preparation and mineral separation. Mr. E.H. Pedersen prepared the thin sections. I wish to thank Mrs. R. Maier in particular for her advice on the chemical processing of samples for uranium-lead. Mr. D.J. Millar provided technical assistance with the mass spectrometry and Mr. R.C. Chamberlain constructed the infracentrifuge.

Dr. G. Woodford devised the method by which the lead data were corrected for contamination and wrote the necessary computer program. Dr. P.A. Arriens was responsible for the programming of the multi-user system to which the mass spectrometers were linked 'on-line'. Considerable assistance was received from Mr. N.W. Tetley in the modification of those programs and the installation of new programs for the acquisition of lead and uranium data. The program used to draw the concordia diagrams was written by Dr. B.W. Chappell.

Dr. J.N. Boland, Messrs P.E. Willis and J. Preston and Mrs. V.
Heaume provided assistance with the photomicrographs.

Dr. W. Compston, Messrs N.W. Tetley, D.A. Nieuwland and R.S. Freeman and Mrs. J.E.H. Williams assisted with the collection of samples.

The script benefited greatly from the constructive criticisms of Drs. W. Compston, R.T. Pidgeon and R.S. Anderssen, Mrs. J.E.H. Williams and Mr. R.I. Hill.

The preparation of the thesis was greatly assisted by the expert typing and advice of Mrs. N. Blundell, Mrs. J. Jenkins and Ms. C.D. Neagle.

The enthusiastic assistance of rangers of the National Parks and Wildlife Service, Mr. J. Gallard and Mr. G. Warboys is gratefully acknowledged.

Thanks are due to Mrs. J.E.H. Williams and Mr. and Mrs. R.I. Hill for their help with proof reading.

I wish to record in addition my deep gratitude to Janet for her invaluable assistance in innumerable ways and to my parents, without whose support over many years this work would never have been done.

Financial support for this research was provided by a Commonwealth Postgraduate Research Award, supplemented by the Australian National University.
ABSTRACT

The Berridale Batholith is one of the numerous Palaeozoic granitic batholiths in southeastern Australia. It is a composite body consisting of more than thirty plutons, in each of which occurs one of the twenty principal granitoids which have been distinguished on petrographic and chemical grounds. The granitoids have been classified as I- or S-type on the basis of whether they have been derived from Igneous or Sedimentary source materials respectively.

A study of the geochronology of selected granitoids by zircon and monazite uranium-lead and by biotite rubidium-strontium, and a comparison of the measured ages with those which other workers have determined by total-rock rubidium-strontium and biotite and hornblende potassium-argon, has revealed significant differences between the results obtained by the different isotopic techniques. These differences are considered to be an indication that a single granitic magma may evolve over an extended period of time.

All the S-type granitoids, and some of the I-type granitoids, contain a zircon component which is older than their oldest possible emplacement ages as determined from the stratigraphic control. Because the granitoids have not assimilated large amounts of country rock, this component is considered in some cases to be zircon which crystallised from the magma at an earlier stage in its history and in other cases to be detrital zircon from the granitoid's source rocks. The minimum provenance age of the oldest zircon in the S-type granitoids, and hence of the zircon in their source material, is 1163 ± 5 m.y. That source material is considered to be metasedimentary basement which is thought to underlie the region.

It is considered that in cases in which a granitoid's zircon discordance pattern is dominated by the effects of inheritance, the geological significance of the intercepts of the discordia with the concordia is
questionable. All the lower intersection ages measured on such samples from the Berridale Batholith are equal to or younger than the ages measured on biotite and monazite from the same samples. For this reason, and because the granitoids' biotite ages are without exception compatible with the intrusive sequence established geologically, the biotite ages are considered to be the most accurate minimum estimates of the plutons' emplacement ages. These ages are in the range $412.3 \pm 1.3$ m.y. to $419.0 \pm 1.8$ m.y., so the plutons are considered to have been emplaced in a short time interval in the Late Silurian to Early Devonian.

The monazite ages are equal to or older than the biotite ages ($412.2 \pm 2.3$ m.y. to $439.8 \pm 3.0$ m.y.) and are considered to be the ages of monazite crystallisation at the time of, or prior to each granitoid's emplacement. The monazite ages are therefore a minimum estimate of the ages of the melting events which formed the granitic magmas. These ages and some of the zircon ages are interpreted as indicating that the magmatic history of many of the plutons extends back at least to the Middle Silurian, and in some cases to the Late Ordovician or earlier.

The granitoids' total-rock rubidium-strontium ages are equal to or older than their biotite ages. The total-rock age indicates the time at which the strontium isotopes in the granitic magma or its source rocks were last equilibrated. The total-rock age may be the age of a melting event or be partially inherited from the granitoid's source material. In either case, the total-rock ages are minimum estimates of the age of the granitoids' source rocks.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Acknowledgements</td>
<td>1</td>
</tr>
<tr>
<td>INTRODUCTION</td>
<td>3</td>
</tr>
<tr>
<td>DECAY CONSTANTS</td>
<td>4</td>
</tr>
<tr>
<td><strong>CHAPTER 1 - THE CONTEXT OF THE PRESENT WORK</strong></td>
<td>4</td>
</tr>
<tr>
<td>1.1 THE STRUCTURE OF THE AUSTRALIAN CONTINENT</td>
<td>4</td>
</tr>
<tr>
<td>1.2 THE TASMAN MOBILE ZONE</td>
<td>8</td>
</tr>
<tr>
<td>1.3 THE BERRIDALE BATHOLITH AND ITS HOST ROCKS</td>
<td>15</td>
</tr>
<tr>
<td>1.3.1 The Adaminaby Beds</td>
<td>10</td>
</tr>
<tr>
<td>1.3.2 The Berridale Batholith</td>
<td>13</td>
</tr>
<tr>
<td>1.4 PREVIOUS STUDIES OF THE GRANITOIDs</td>
<td>15</td>
</tr>
<tr>
<td>1.4.1 Early work</td>
<td>18</td>
</tr>
<tr>
<td>1.4.2 I- and S-type granitoids</td>
<td>21</td>
</tr>
<tr>
<td>1.4.3 Isotopic studies</td>
<td>27</td>
</tr>
<tr>
<td>1.4.3.1 Radiometric isotopes</td>
<td>27</td>
</tr>
<tr>
<td>1.4.3.2 Stable isotopes</td>
<td>27</td>
</tr>
<tr>
<td>1.5 GRANITOID GENESIS</td>
<td>28</td>
</tr>
<tr>
<td>1.6 THE OBJECTIVES OF THE PRESENT STUDY</td>
<td>29</td>
</tr>
<tr>
<td><strong>CHAPTER 2 - SELECTION OF SAMPLES</strong></td>
<td>33</td>
</tr>
<tr>
<td>SUMMARY</td>
<td>40</td>
</tr>
<tr>
<td><strong>CHAPTER 3 - THE MAFFRA ADAMELLITE</strong></td>
<td>41</td>
</tr>
<tr>
<td>3.1 THE GRANITOID</td>
<td>41</td>
</tr>
<tr>
<td>3.2 THE ZIRCON</td>
<td>43</td>
</tr>
<tr>
<td>3.3 GEOCHRONOLOGY</td>
<td>44</td>
</tr>
<tr>
<td>3.3.1 Monazite</td>
<td>48</td>
</tr>
<tr>
<td>3.3.2 Zircon</td>
<td>57</td>
</tr>
<tr>
<td>3.4 SUMMARY</td>
<td>57</td>
</tr>
<tr>
<td><strong>CHAPTER 4 - THE DELEGATE ADAMELLITE</strong></td>
<td>58</td>
</tr>
<tr>
<td>4.1 THE GRANITOID</td>
<td>58</td>
</tr>
<tr>
<td>4.2 THE ZIRCON</td>
<td>60</td>
</tr>
<tr>
<td>4.3 GEOCHRONOLOGY</td>
<td>60</td>
</tr>
<tr>
<td>4.3.1 Rubidium–strontium and potassium–argon analyses</td>
<td>60</td>
</tr>
<tr>
<td>4.3.2 Uranium–lead analyses</td>
<td>61</td>
</tr>
<tr>
<td>4.3.3 The Delegate Aplite</td>
<td>65</td>
</tr>
<tr>
<td>4.3.4 The stratigraphic time scale</td>
<td>68</td>
</tr>
<tr>
<td>4.4 INTERPRETATION OF THE GEOCHRONOLOGY</td>
<td>74</td>
</tr>
<tr>
<td>4.4.1 A geochronological model</td>
<td>76</td>
</tr>
<tr>
<td>4.5 SUMMARY</td>
<td>77</td>
</tr>
<tr>
<td><strong>CHAPTER 5 - THE TARA GRANODIORITE</strong></td>
<td>78</td>
</tr>
<tr>
<td>5.1 THE GRANITOID</td>
<td>78</td>
</tr>
<tr>
<td>5.2 THE ZIRCON</td>
<td>79</td>
</tr>
<tr>
<td>5.3 GEOCHRONOLOGY</td>
<td>81</td>
</tr>
<tr>
<td>5.3.1 Rubidium–strontium and potassium–argon analyses</td>
<td>81</td>
</tr>
<tr>
<td>5.3.2 Uranium–lead analyses</td>
<td>83</td>
</tr>
<tr>
<td>5.3.2.1 The systematics of discordance</td>
<td>83</td>
</tr>
<tr>
<td>5.4 THE TARA XENOLITH</td>
<td>92</td>
</tr>
<tr>
<td>Section</td>
<td>Page</td>
</tr>
<tr>
<td>---------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>5.5 Zircon in the Tara Xenolith</td>
<td>93</td>
</tr>
<tr>
<td>5.6 Uranium-Lead Analyses from the Tara Xenolith</td>
<td>93</td>
</tr>
<tr>
<td>5.7 Interpretation of the Geochronology</td>
<td>97</td>
</tr>
<tr>
<td>5.7.1 A geochronological model</td>
<td>99</td>
</tr>
<tr>
<td>5.8 Summary</td>
<td>100</td>
</tr>
</tbody>
</table>

**CHAPTER 6 - THE S-TYPE GRANITOIDs**

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.1 The Cootralantra Granodiorite</td>
<td>103</td>
</tr>
<tr>
<td>6.1.1 The granitoid</td>
<td>103</td>
</tr>
<tr>
<td>6.1.2 The zircon</td>
<td>104</td>
</tr>
<tr>
<td>6.1.3 Geochronology</td>
<td>106</td>
</tr>
<tr>
<td>6.1.3.1 Rubidium-strontium and potassium-argon analyses</td>
<td>106</td>
</tr>
<tr>
<td>6.1.3.2 Uranium-lead analyses</td>
<td>107</td>
</tr>
<tr>
<td>6.1.4 The Cootralantra xenolith</td>
<td>111</td>
</tr>
<tr>
<td>6.1.5 Zircon in the Cootralantra xenolith</td>
<td>111</td>
</tr>
<tr>
<td>6.1.6 Uranium-lead analyses from the Cootralantra xenolith</td>
<td>113</td>
</tr>
<tr>
<td>6.2 The Bullenbalong Granodiorite</td>
<td>117</td>
</tr>
<tr>
<td>6.2.1 The granitoid</td>
<td>117</td>
</tr>
<tr>
<td>6.2.2 The zircon</td>
<td>117</td>
</tr>
<tr>
<td>6.2.3 Geochronology</td>
<td>119</td>
</tr>
<tr>
<td>6.2.3.1 Rubidium-strontium analyses</td>
<td>119</td>
</tr>
<tr>
<td>6.2.3.2 Uranium-lead analyses</td>
<td>120</td>
</tr>
<tr>
<td>6.3 The Numbla Vale Adamellite</td>
<td>120</td>
</tr>
<tr>
<td>6.3.1 The granitoid</td>
<td>120</td>
</tr>
<tr>
<td>6.3.2 The zircon</td>
<td>124</td>
</tr>
<tr>
<td>6.3.3 Geochronology</td>
<td>124</td>
</tr>
<tr>
<td>6.3.3.1 Rubidium-strontium and potassium-argon analyses</td>
<td>124</td>
</tr>
<tr>
<td>6.3.3.2 Uranium-lead analyses</td>
<td>126</td>
</tr>
<tr>
<td>6.4 The Dalgety Granodiorite</td>
<td>129</td>
</tr>
<tr>
<td>6.4.1 The granitoid</td>
<td>129</td>
</tr>
<tr>
<td>6.4.2 The zircon</td>
<td>130</td>
</tr>
<tr>
<td>6.4.3 Geochronology</td>
<td>131</td>
</tr>
<tr>
<td>6.4.3.1 Rubidium-strontium and potassium-argon analyses</td>
<td>131</td>
</tr>
<tr>
<td>6.4.3.2 Uranium-lead analyses</td>
<td>133</td>
</tr>
<tr>
<td>6.5 The Bog Hole Adamellite</td>
<td>136</td>
</tr>
<tr>
<td>6.5.1 The granitoid</td>
<td>136</td>
</tr>
<tr>
<td>6.5.2 The zircon</td>
<td>137</td>
</tr>
<tr>
<td>6.5.3 Geochronology</td>
<td>137</td>
</tr>
<tr>
<td>6.5.3.1 Uranium-lead analyses</td>
<td>137</td>
</tr>
<tr>
<td>6.6 Interpretation of the Geochronology</td>
<td>141</td>
</tr>
<tr>
<td>6.6.1 The Dalgety Sandstone</td>
<td>147</td>
</tr>
<tr>
<td>6.7 Summary</td>
<td>155</td>
</tr>
</tbody>
</table>

**CHAPTER 7 - THE FINISTER GRANITOID**

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.1 The Granitoid</td>
<td>157</td>
</tr>
<tr>
<td>7.2 The Zircon</td>
<td>158</td>
</tr>
<tr>
<td>7.3 Geochronology</td>
<td>160</td>
</tr>
<tr>
<td>7.3.1 Rubidium-strontium and potassium-argon analyses</td>
<td>161</td>
</tr>
<tr>
<td>7.3.2 Uranium-lead analyses</td>
<td>165</td>
</tr>
<tr>
<td>7.4 Interpretation of the Geochronology</td>
<td>168</td>
</tr>
<tr>
<td>7.4.1 A geochronological model</td>
<td>171</td>
</tr>
</tbody>
</table>
CHAPTER 8 - THE GEOCHRONOLOGY OF THE BERRIDALE BATHOLITH

APPENDIX A - ANALYTICAL TECHNIQUES

A.1 URANIUM-LEAD FROM ZIRCON AND MONAZITE
   A.1.1 Crushing
   A.1.2 Mineral separation
   A.1.3 Chemical processing
      A.1.3.1 Dissolution
      A.1.3.2 Spiking
      A.1.3.3 Extraction of uranium and lead

A.2 RUBIDIUM-STRONTIUM
   A.2.1 Crushing
   A.2.2 Mineral separation
   A.2.3 Chemical processing

A.3 SAMPLE LOADING FOR MASS SPECTROMETRY

APPENDIX B - DATA REDUCTION

B.1 THE 'ON-LINE' PROCESSING OF MASS SPECTROMETER DATA
   B.1.1 Uranium data
   B.1.2 Lead data

B.2 THE CORRECTION OF ISOTOPIC ANALYSES FOR FRACTIONATION AND CONTAMINATION
   B.2.1 An outline of the mathematical solution

APPENDIX C - ANALYTICAL PRECISION AND ACCURACY

C.1 CALIBRATION OF THE MSZ MASS SPECTROMETER
   C.1.1 Electrometer response
   C.1.2 The measurement of small signals

C.2 ANALYSES OF STANDARDS
   C.2.1 Rubidium-strontium standards
      C.2.1.1 SRM 987 strontium standard
      C.2.1.2 SRM 607 K-feldspar standard
   C.2.2 Uranium-lead standard
      C.2.2.1 SRM 981 common lead standard
      C.2.2.2 SRM 983 radiogenic lead standard
      C.2.2.3 SRM 613 trace elements in a glass matrix
      C.2.2.4 GA834-A zircon standard

C.3 THE ASSESSMENT OF ANALYTICAL ERROR
   C.3.1 Error in the isotopic analysis of lead
      C.3.1.1 Error in $^{206}$Pb*
      C.3.1.2 Error in $^{207}$Pb*
   C.3.2 Error in the lead-uranium ratio

APPENDIX D - ANALYSIS OF VARIANCE FOR THE RUBIDIUM-STRONTIUM BIOTITE AGES

APPENDIX E - URANIUM-LEAD ANALYSES OF POTASSIUM FELDSPAR

E.1 CRUSHING AND MINERAL SEPARATION
E.2 CHEMICAL PROCESSING

APPENDIX F - SAMPLE LOCALITIES

F.1 ZIRCON SAMPLES
FIGURES

1.1 The main basement provinces of the Australian continent. 5
1.2 A sketch map of the granitoids of southeastern Australia. 9
1.3 Granitoids of the Berridale and Kosciusko Batholiths. 11
2.1 Intrusion relationships in the Berridale Batholith. 34
3.1 The distribution of zircon in the minerals of the Maffra Adamellite. 42
3.2 Concordia plot of the uranium-lead data from the Maffra Adamellite. 46
3.3 The effect on the Maffra uranium-lead data of increasing the common lead $^{207}$Pb/$^{206}$Pb. 50
3.4 Inherited lead in fraction 43/. 52
4.1 The distribution of zircon in the minerals of the Delegate Adamellite. 59
4.2 Concordia plot of the uranium-lead data from the Delegate Adamellite. 64
4.3 Isochron plot for the Delegate Aplite. 67
4.4 Isochron plot for the Kinnekuile A bentone. 72
5.1 The distribution of zircon in the minerals of the Tara Granodiorite. 80
5.2 Concordia plot of the uranium-lead data from the Tara Granodiorite. 85
5.3 Tara Granodiorite zircon, $^{206}$Pb/$^{238}$U age vs. uranium concentration. 88
5.4 Tara Granodiorite zircon, uranium concentration vs. grain size. 89
5.5 Tara Granodiorite zircon, $^{207}$Pb/$^{206}$Pb age vs. grain size. 90
5.6 Tara Granodiorite zircon, $^{207}$Pb/$^{206}$Pb age vs. uranium concentration. 91
5.7 The distribution of zircon in the minerals of the Tara xenolith. 94
5.8 Concordia plot of the uranium-lead data from the Tara xenolith. 96
5.9 Concordia plot of the combined uranium-lead data from the Tara xenolith and Tara granitoid. 98
6.1 The distribution of zircon in the minerals of the Costralantra Granodiorite. 105
6.2 Concordia plot of the uranium-lead data from the Cootralantra Granodiorite.

6.3 The distribution of zircon in the minerals of the Cootralantra xenolith.

6.4 Concordia plot of the uranium-lead data from the Cootralantra xenolith.

6.5 Concordia plot of the combined uranium-lead data from the Cootralantra xenolith and Cootralantra granitoid.

6.6 The distribution of zircon in the minerals of the Bullenbalong Granodiorite.

6.7 Concordia plot of the uranium-lead data from the Bullenbalong Granodiorite.

6.8 The distribution of zircon in the minerals of the Numbla Vale Adamellite.

6.9 Concordia plot of the uranium-lead data from the Numbla Vale Adamellite.

6.10 The distribution of zircon in the minerals of the Dalgety Granodiorite.

6.11 Concordia plot of the uranium-lead data from the Dalgety Granodiorite.

6.12 The distribution of zircon in the minerals of the Bog Hole Granite.

6.13 Concordia plot of the uranium-lead data from the Bog Hole Granite.

6.14 Inherited lead in fraction 79/.

6.15 Concordia plot of the uranium-lead data from the Dalgety Sandstone.

7.1 The distribution of zircon in the minerals of the Finister Granodiorite.

7.2 Comparison of biotite rubidium-strontium and potassium-argon ages from the Finister Granodiorite.

7.3 Concordia plot of the uranium-lead data from the Finister Granodiorite.

7.4 Finister Granodiorite zircon, $^{207}\text{Pb}/^{206}\text{Pb}$ age vs. uranium concentration.

A.1 Principal components of the continuous-feed, overflow infracentrifuge.

C.1A Dynamic zero vs. reference signal, June, 1975, dependence upon range.

C.1B Dynamic zero vs. reference signal, May, 1975, dependence upon delay.
C.2A Dynamic zero vs. reference signal, October, 1975. 211
C.2B Dynamic zero vs. reference signal, October, 1976. 211
C.3A A dynamic zero vs. delay time. 213
C.3B Decay of residual signal with time. 213
C.4 Δ vs. input voltage, June, 1976. 217
C.5 SRM981 common lead standard, 207Pb/206Pb vs. 208Pb/206Pb. 221
C.6 SRM983 radiogenic lead standard, 207Pb/206Pb vs. 208Pb/206Pb. 223
C.7 SRM983 radiogenic lead standard, 204Pb/206Pb vs. 208Pb/206Pb. 224
C.8 Concordia plot of the uranium-lead data from the Cooma Syenite. 229
E.1 K-feldspar from the Berridale Batholith, 207Pb/204Pb vs. 206Pb/204Pb. 256
F.1 The granitoids and zircon sample localities in the Berridale and Kosciusko Batholiths. 261
F.2 The granitoids and biotite sample localities in the Berridale Batholith. 262
<table>
<thead>
<tr>
<th>Table</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1</td>
<td>Uranium-lead analytical data from the Maffra Adamellite (75-117).</td>
<td>45</td>
</tr>
<tr>
<td>4.1</td>
<td>Uranium-lead analytical data from the Delegate Adamellite (75-120).</td>
<td>63</td>
</tr>
<tr>
<td>4.2</td>
<td>Rubidium, strontium and $^{87}\text{Sr}/^{86}\text{Sr}$ for the Delegate Aplite.</td>
<td>66</td>
</tr>
<tr>
<td>4.3</td>
<td>Rubidium, strontium and $^{87}\text{Sr}/^{86}\text{Sr}$ for the Kinnekulle A.</td>
<td>71</td>
</tr>
<tr>
<td>4.8</td>
<td>$^{87}\text{Sr}/^{86}\text{Sr}$ for the Delegate Aplite.</td>
<td>66</td>
</tr>
<tr>
<td>5.1</td>
<td>Uranium-lead analytical data from the Tara Granodiorite (75-002).</td>
<td>84</td>
</tr>
<tr>
<td>5.2</td>
<td>Uranium-lead analytical data from the Tara xenolith (77-038).</td>
<td>95</td>
</tr>
<tr>
<td>6.1</td>
<td>Total-rock model ages for the Cootralantra Granodiorite.</td>
<td>107</td>
</tr>
<tr>
<td>6.2</td>
<td>Uranium-lead analytical data from the Cootralantra Granodiorite (75-109).</td>
<td>108</td>
</tr>
<tr>
<td>6.3</td>
<td>Uranium-lead analytical data from the Cootralantra xenolith (77-039).</td>
<td>114</td>
</tr>
<tr>
<td>6.4</td>
<td>Uranium-lead analytical data from the Bullenbalong Granodiorite (77-264).</td>
<td>121</td>
</tr>
<tr>
<td>6.5</td>
<td>Uranium-lead analytical data from the Numbla Vale Adamellite (75-107).</td>
<td>127</td>
</tr>
<tr>
<td>6.6</td>
<td>Uranium-lead analytical data from the Dalgety Granodiorite (75-113).</td>
<td>134</td>
</tr>
<tr>
<td>6.7</td>
<td>Uranium-lead analytical data from the Bog Hole Granite (77-040).</td>
<td>139</td>
</tr>
<tr>
<td>6.8</td>
<td>Uranium-lead analytical data from the Dalgety Sandstone (76-368).</td>
<td>149</td>
</tr>
<tr>
<td>7.1</td>
<td>Total-rock model ages for the Finister Granodiorite.</td>
<td>162</td>
</tr>
<tr>
<td>7.2</td>
<td>Uranium-lead analytical data from the Finister Granodiorite (74-335).</td>
<td>166</td>
</tr>
<tr>
<td>B.1</td>
<td>Table of t.</td>
<td>199</td>
</tr>
<tr>
<td>C.1</td>
<td>Linearity calibrations.</td>
<td>216</td>
</tr>
<tr>
<td>C.2</td>
<td>Rubidium and strontium analyses of SRM607.</td>
<td>219</td>
</tr>
<tr>
<td>C.3</td>
<td>Lead analyses of SRM981.</td>
<td>220</td>
</tr>
<tr>
<td>C.4</td>
<td>Lead analyses of SRM983.</td>
<td>222</td>
</tr>
<tr>
<td>C.5</td>
<td>Calibration of lead and uranium in the $^{208}\text{Pb}/^{235}\text{U}$ tracer against SRM613.</td>
<td>226</td>
</tr>
<tr>
<td>C.6</td>
<td>Uranium-lead analytical data from the Cooma Syenite (GA834-A).</td>
<td>228</td>
</tr>
</tbody>
</table>
D.1 Analytical data for biotites from the Berridale Batholith.
D.2 Computations for nested classification analysis of variance.
D.3 Analysis of variance.
D.4 Significance tests.
D.5 Analysis of variance.
D.6 Significance tests.
E.1 Analytical data for Berridale Batholith K-feldspar.
E.2 Common lead compositions.
INTRODUCTION

This thesis reports the results of research into aspects of the lead and strontium geochronology of the granitoids of the Berridale Batholith (southeastern Australia), which was undertaken at the Research School of Earth Sciences, Australian National University, between January 1974 and September 1977. The work reported is part of a larger project initiated several years previously by Drs. A.J.R. White, B.W. Chappell and W. Compston, a combined geological, geochemical and isotopic study of the extensive mid-Palaeozoic plutonism in southeastern Australia.

The principal aim of the present research was to determine the age and prehistory of selected granitoids by the analysis of the lead-uranium system in zircon. In practice, these analyses were augmented by others of the uranium-lead system in monazite and the rubidium-strontium system in total-rocks and biotite. Because this work was the first major isotopic study of zircon to be done at the Research School, a significant proportion of the thesis has been devoted to a discussion of analytical techniques, mass spectrometer calibrations and data reduction. Each of these was an important factor in determining the accuracy of the final data.

The thesis has three principal sections. The first is primarily introductory and places the present study within the context of the work done by other researchers. The interpretation of the isotopic data rested heavily upon that which already was known about the granitoids: their geological relationships, their geochemistry and their place in the geological history both of the Snowy Mountains region and of the Australian continent as a whole. The second is a presentation and discussion of the analytical data. The third is a set of appendices which deal with technical aspects of the study: analytical techniques, calibrations and data reduction. Included in this section are details of the samples which were analysed.
The decay constants used in the calculation of isotopic ages are listed in the section below. Unless stated otherwise, the estimates of the precision of the age determinations are quoted as two standard errors.
DECAY CONSTANTS

Unless stated otherwise, the isotopic ages in this thesis have been calculated using the standard set of decay constants and isotopic abundances recommended by the Subcommission on Geochronology of the IUGS Commission of Stratigraphy (Steiger & Jäger, 1976).

\[ \lambda^{238}\text{U} = 1.55125 \times 10^{-10} \text{ yr}^{-1} \]

\[ \lambda^{235}\text{U} = 9.8485 \times 10^{-10} \text{ yr}^{-1} \]

(Jaffey et al., 1971)

\[ \frac{^{238}\text{U}}{^{235}\text{U}} = 137.88 \]

(Shields, 1960; Cowan & Adler, 1976)

\[ \lambda^{87}\text{Rb} = 1.42 \times 10^{-11} \text{ yr}^{-1} \]

(recommended by the Subcommission for provisional use)

\[ \frac{^{88}\text{Sr}}{^{86}\text{Sr}} = 8.3752 \]

(Nier, 1938)

\[ \lambda^{40}\text{K}^- = 4.962 \times 10^{-10} \text{ yr}^{-1} \]

\[ \lambda^{40}\text{K}^- + \lambda^{40}\text{K}^+ = 0.581 \times 10^{-10} \text{ yr}^{-1} \]

(Beckinsale & Gale, 1969)

\[ ^{39}\text{K} = 93.2581 \text{ atom percent} \]

\[ ^{40}\text{K} = 0.01167 \text{ atom percent} \]

\[ ^{41}\text{K} = 6.7302 \text{ atom percent} \]

(Garner et al., 1976)

In accordance with the original value used in spike calibrations \( ^{85}\text{Rb}/^{87}\text{Rb} \) was assumed to be 2.600 for samples analysed by isotope dilution at Earth Sciences.
CHAPTER 1

THE CONTEXT OF THE PRESENT WORK

1.1 THE STRUCTURE OF THE AUSTRALIAN CONTINENT

As with each of the world's large landmasses, the Australian continent is composed of several crustal segments, each with an unique tectonic history. These may be grouped so as to divide the continent into two parts, the Precambrian basement provinces which form the shield and the Palaeozoic basement provinces which comprise the Tasman Mobile Zone.

Rutland (1976) considered the main components of the shield to be three blocks, the Yilgarn, the Pilbara and the Kimberley, and two composite provinces, the Arunta and the Gawler (Fig. 1.1). Other authors have attempted a subdivision of the Arunta and Gawler (for example, Plumb, 1976) but this has been made difficult by extensive early Palaeozoic platform cover.

Geochronological studies reported by Compston and Arriens (1968) have provided a measure of the time interval over which the shield evolved. The oldest rocks have been found in the Yilgarn and Pilbara blocks, some of the ages being of the order of 3.0 b.y. Elsewhere, ages in access of 2.0 b.y. are uncommon. The significant period was 1.6 to 1.9 b.y., during which widespread sedimentation was initiated. Tectonism and sedimentation continued until about 1.0 b.y. ago, by which time the shield had attained essentially a stable configuration. The southeastern edge of the Gawler province, however, was active as late as the Early Cambrian. It was against this stable Precambrian block that the Tasman Mobile Zone evolved.

1.2 THE TASMAN MOBILE ZONE

The pre-Ordovician history of the Tasman Mobile Zone largely remains unresolved. Development of the Zone on oceanic crust (Crook, 1969, 1974), continental crust (Hills, 1956; Rutland, 1973, 1976) or some combination of the two (Scheibner, 1974, 1976) has been postulated. The geophysical data which presently are available do not provide a unique solution to the problem but do suggest that the shield as it is now recognised does no
Fig. 1.1 The main basement provinces of Australia (after Rutland, 1976). The smaller sub-provinces of the Arunta, Gawler and Tasman are not shown. The approximate easterly limit of Precambrian rocks discussed by Howard & Sass (1964) is marked by the line P-P.
extend much further to the east than its easternmost surface outcrop.

In a study of P wave travel times along a traverse from central South Australia to the eastern coast, Cleary et al. (1972) obtained results consistent with 'a wedging out of the lower velocity layer in the upper mantle as the shield boundary is approached' (p.4). They found the most rapid westerly decrease in P station anomalies to coincide with the eastern boundary of the Precambrian proposed by Howard and Sass (1964) (Fig. 1.1). Wellman (1976) placed the eastern extreme of his 'area of Proterozoic stabilisation' in a similar position, having used gravity trends in the definition of Australia's principal crustal units.

Indirect evidence for at least a portion of the Tasman Mobile Zone being underlain by pre-Ordovician continental crust has been advanced by White et al. (1976). They considered the areal distribution of mid-Palaeozoic granitoids of fundamentally different chemistries to imply the presence of metasedimentary basement beneath the Snowy Mountains region at the southeastern end of the Zone. Seismic studies indicate a crustal thickness of about 42 kilometres in the region (Doyle et al., 1966) but this may well be only an isostatic effect, since the crust thins rapidly further to the west (Muirhead et al., 1977). The interpretation of the gravity data for the region is complicated by the presence of the mountains and the nearness of the coast.

There have been numerous attempts made to apply the plate tectonic model to the Tasman Mobile Zone (Oversby, 1971, 1972; Solomon and Griffiths, 1972; Scheibner, 1972a,b; Packham, 1973 and others) but a consensus has yet to be reached on the location of the major tectonic units (Packham & Leitch, 1974). Most interpretations have in common, however, the concept of the buildup of the Mobile Zone eastward from the stabilised Precambrian shield by the accretion of a succession of island arcs and basins.

Brown et al. (1968) considered the geological evolution of the Mobile Zone through an interpretation of the stratigraphic record. Although
a considerable amount of additional information (both published and unpublished) has become available since then, their work remains a far less interpretative study of the geology than any one of the variety of regional syntheses of later authors. Brown et al.'s (1968) discussion of the southeastern end of the Mobile Zone (called at that time the 'Lachlan Geosyncline' but now the 'Lachlan Mobile Zone') may be generalised as follows.

The oldest rocks in the Lachlan Zone were thought to be Cambrian, although no fossils of that age had then, or have since, been found. The rocks in question are a thick sequence of low-grade metamorphics which lie unconformably below Late Ordovician rocks along the southeastern coast and in the Snowy Mountains region.

The stratigraphic record was much more comprehensive for the Ordovician. Several meridional troughs and highs developed. The troughs received a thick sequence of quartz-rich sediments (some of which are richly graptolitic) ranging from deep-water greywacke, shale and chert to shallow-water sandstone and some volcanics. On the highs, shallow-water sediments (for example greywacke, sandstone and limestone) and volcanics accumulated. The relative tectonic stability of the period ended in the Late Ordovician with the Benambran Orogeny, during which the trough-and-arch pattern was reorganised and there was intense local deformation. Felsic and intermediate volcanism which followed was largely restricted to the deformed areas.

The Silurian was a period of relative stability. Greywacke-shale deposition continued in a trough on the western side of the Lachlan Zone while the areas that had been uplifted by the Benambran event accumulated shallow-water sediments (sandstone, limestone and shale), interspersed with felsic to intermediate volcanics. Towards the end of the period there began a tectonic episode (the Bowning Orogeny) which continued on into the Early Devonian. Much of the eastern half of the Zone was uplifted and there, no further deposition took place until the Late Devonian.

The Bowning event corresponded with the period of the most widespread plutonism in the Lachlan Zone. The plutonics are broadly
co-extensive with the distribution of the earlier felsic volcanism. If a genetic link between the two is assumed, they represent a protracted igneous episode which extended from the Llandoverian to the Early Devonian. It was during this period that the granitoids of the Berridale Batholith were emplaced.

The Bowning Orogeny caused widespread uplift in most parts of the Mobile Zone. Although deposition continued without interruption in the trough on the western side of the Zone, through the Early and Middle Devonian it became a shallow water basin. The remainder of the Zone became either an area of shallow-water sedimentation or rose above sealevel. Volcanism, (some intermediate but predominantly felsic) was widespread. The Middle Devonian Tabberabberan Orogeny disrupted completely the trough-and-arch pattern and virtually the whole Zone was raised above sea level.

Following the Tabberabberan event, southeastern Australia behaved essentially as a stable continental block. Igneous activity has continued into historical times (Carboniferous to Permian felsic volcanism and plutonism in the north, Permian basic volcanism and scattered Jurassic acid plutonism on the southeastern coast and Tertiary to recent basic volcanism in the south and southwest) but nowhere has it been accompanied by major tectonism. The post-Devonian cover was deposited principally during Permian to Jurassic and Tertiary marine transgressions and has been augmented by Quaternary terrestrial deposits.

1.3 THE BERRIDALE BATHOLITH AND ITS HOST ROCKS

The Late Silurian to Early Carboniferous plutonism in the Lachlan Zone was so extensive that, of the present area of exposed Palaeozoic rock, nearly twenty-five percent is accounted for by granitic batholiths (Fig. 1.2). The term 'batholith' is used here to describe a composite body of granitic rock several hundred square kilometres in area which is isolated from other similar bodies by intervening country rock. Although this inherently implies a close spatial relationship between the constituent intrusives of a batholith
Fig. 1.2 A sketch map of southeastern Australia showing the principal granitoids of the Lachlan Mobile Zone and the location of the Berridale and Kosciusko Batholiths.
a close genetic relationship does not necessarily follow.

The Berridale Batholith (Fig. 1.2, 1.3) was emplaced near the southern end of the Lachlan Mobile Zone into one area that is now a tableland on the eastern flank of the Snowy Mountains. The host rock to the Batholith is a thick sequence of quartz-rich greywacke, siltstone, sandstone and shale, interbedded with lesser amounts of black slate and chert. Relatively poor exposure, the limited distribution of fossils, locally intense folding and disruption of the sequence by the granitoids has made systematic work on the host rock particularly difficult.

The stratigraphy at the northern end of the Batholith has been reviewed by White et al. (1977) and their conclusions are equally applicable to the sequence further south. It was their opinion that, despite several local studies in which stratigraphic units had been delineated (for example Joplin, 1942; Adamson, 1955; Fairbridge et al., 1951 in Moye et al., 1969 and Crook et al., 1973), the difficulties of correlation beyond the type areas were considerable and no satisfactory subdivision of the sediments had been made. As a consequence, White et al. (1977) used a single name for all the flysch-facies sedimentary rocks of the area, the 'Adaminaby Beds'.

1.3.1 The Adaminaby Beds

In the course of geological mapping for the present and associated studies (White et al., 1976, 1977, 1978), a considerable amount of descriptive data on the Adaminaby Beds has been accumulated. This is insufficient, however, to permit the construction of a formalised stratigraphy.

The Adaminaby Beds are thought to be predominantly Late Ordovician in age. Richly graptolitic black slate has been found at numerous localities and the fact that, with few exceptions, the graptolite populations are Eastonian, indicates either voluminous sedimentation at that time or structural complexities which have exposed a small number of Eastonian horizons repeatedly over a large area. The exceptions are localities of slightly older (Gisbornian) or slightly younger (Bolindian) assemblages
Fig. 1.3 A geological map of the Berridale Batholith and the eastern side of the Kosciusko Batholith showing the distribution of I- and S-type granitoids (White et al., 1977, 1978). The principal faults in the region are a set of conjugate wrench faults and meridional thrusts.
Pre-Eastonian rocks are best exposed near the southern end of the Batholith. There, several hundred metres of thick-bedded sandstone underlie graptolitic black slate. Sandstone which is similar occurs further to the north and is the immediate host to some of the intrusives but its relationship to fossiliferous units has not been established directly.

On the western side of the Batholith, southeast of Jindabyne (Fig. 1.3), a simple sequence has been mapped which includes two graptolitic units, one of Eastonian and one of Bolindian age. The pre-Eastonian sequence largely has been intruded out but as much as 1000 metres of sediment overlies the Bolindian strata. The possibility that some of that sediment is Silurian currently is being investigated.

Deformation of the Adaminaby Beds is a regional feature which bears no obvious relationship to the Batholith. Most structures in the country rock which were traced to the Batholith margin were found to be truncated by the intrusives. An exception was a unit of felsic volcanics (the Bennetts Creek Volcanics) which appears to have been displaced and refolded by one of the plutons (White et al., 1977). In broad terms, the strength of the regional deformation increases from the south to the north.

In the south, the folding is open and relatively simple. The major structures may be delineated using markers such as black slate horizons. Fifty kilometres further to the north, west of the central region of the Batholith (the eastern side is overlain by Tertiary basalt) the strike of the bedding and fold axes is consistently close to meridional and a cleavage has developed parallel to the bedding planes. The folding is open nevertheless, and it has been possible to establish simple successions such as the one described above.

The deformation in the country rock at the northern end of the Batholith is considerably stronger. Bedding and cleavage coincide, are near to vertical and strike approximately meridionally over large areas. Marker beds such as black slate and the Bennetts Creek Volcanics may be
traced for tens of kilometres parallel to strike before they wedge out. On these grounds and from the presence of a recognisable recumbent fold in one small area of lesser deformation, White et al. (1977) concluded that the area is one 'of tightly appressed isoclinal folds and large scale superposition'.

Unlike the nearby Binjura Beds (Joplin, 1942), the Adaminaby Beds do not show any effects of regional metamorphism. The only rocks in the Adaminaby Beds which are metamorphosed are those within the contact aureole of an intrusive. The aureoles, defined by the limit of spotting in pelitic hand specimens, rarely exceed 1 kilometre in width. Commonly, it is only within about 300 metres of the granitoid-sediment contact that hardened hornfels has developed. In several places recognisable graptolites have been found within rocks of the hornfels zone, attesting to the lack of internal strain in the country rock as the Batholith was emplaced.

1.3.2 The Berridale Batholith

The Berridale Batholith (Fig. 1.3), having an outcrop area of approximately 1700 square kilometres, is one of the smallest batholiths in the Lachlan Mobile Zone. It is a composite body and includes twenty major granitoids in the compositional range tonalite to adamellite and numerous small aplitic intrusives. The granitoid nomenclature used here is that recommended by the IUGS Subcommission on the Systematics of Igneous Rocks but, in line with common usage in Australia, the term 'Adamellite' is retained to describe granitoids with alkali feldspar between 35 and 65% of the total feldspar.

A very large part of the Berridale Batholith is accounted for by granodiorites and adamellites, and tonalites are a very minor constituent. Each granitoid (which has been given as stratigraphic name for ease of reference) occurs in a discrete intrusive unit which, in plan, most commonly is a quasi-ellipse or a complex of the same. Each elliptical body is considered to be a pluton emplaced during a single intrusive event. The
absence of topographic control on the pluton boundaries implies that they are near to vertical, which in turn suggests that a close approximation to a pluton's three dimensional form is a vertical 'elliptical cylinder'.

The Batholith and the plutons in it are elongated approximately meridionally, parallel to the structural grain which pervades the other rocks of the southeastern end of the Lachlan Zone. This is not considered to be the consequence of post-intrusive deformation but rather a reflection of the presence of a regional stress field at the time the granitoids were emplaced (White et al., 1974).

The region of the Berridale Batholith is transected by numerous inter-related faults (Fig. 1.3). The set developed most extensively is a conjugate system of horizontal wrench faults, in which the displacement on those striking northeasterly is dextral and on those striking northwesterly is sinistral. Bisecting the angle between the wrenches, as does the direction of elongation of the Batholith, are several northward-striking thrusts. Overthrusting has occurred both towards the east and the west. The implication of this fault pattern is that the region has been (and may still be) subjected to considerable east-west compression.

During the emplacement of the Batholith, the direction of principal stress was similarly east-west. Many aplite dykes were intruded which are aligned parallel to either the present wrench or thrust fault directions (White et al., 1974, 1976, 1977), suggesting that these reflect the orientation of the planes of greatest weakness at the time of intrusion. Less commonly, a similar effect is seen in oriented protrusions from otherwise elliptical plutons (White et al., 1974).

The plutons have little internal structure; each granitoid is characterised by a limited range in texture and mineralogy so most of the plutons are outwardly almost homogeneous. There are no examples of major concentric structures such as compositional zonation or ring faulting and primary foliation is very rare. Some of the granitoids have a weak secondary foliation defined by a preferential alignment of the minerals,
especially the micas. The foliation varies in strength and orientation on a regional basis as well as being developed more strongly in some granitoids than in others within a particular area. This, and the facts that the foliation traverses pluton boundaries without deviation and is sub-parallel to the Batholith's elongation, is further support for the postulate of regional directed stress at, or soon after, the time of emplacement.

1.4 PREVIOUS STUDIES OF THE GRANITOIDS

1.4.1 Early work

Although the most rapid advancement of understanding of the Lachlan Zone granitoids has been in the last fifteen years, the history of researchers' continuing interest in the southern batholiths began many years beforehand. The earliest studies were, of necessity, undertaken on a reconnaissance basis and concentrated on the delineation of the major rock units. For example Browne (1929), in his resume of the igneous history of New South Wales, made no mention of the composite nature of the batholiths beyond noting that the rocks of the Kosciusko Batholith (Fig. 1.2) contained a superimposed gneissic foliation which varied in strength from place to place. This particular feature was to play an important rôle as a criterion for granitoid classification in the schemes of later workers.

The main impetus to geological studies in the Snowy Mountains region was the planning of the Snowy Mountains Hydro-electric Scheme in the late 1940s. The investigations of proposed tunnel lines and dam sites by officers of the Bureau of Mineral Resources, Geology and Geophysics led to the compilation of some the first detailed geological maps of the area. The planning of the Scheme also resulted in the initiation of a more wideranging program of reconnaissance mapping.

From the detailed mapping came the first petrologically-based classification of the granitoids. W.B. Dallwitz subdivided the granitoids which occurred along a tunnel line west of Jindabyne into 'massive' and
'foliated' and proposed that the differences in texture were closely related to differences in mineralogy (Ball et al., 1948). Dallwitz argued that the differences reflected different modes of origin, the large 'foliated' granitoids being the product of \textit{in situ} granitisation of sediments and the smaller, later 'massive' granitoids representing remobilised fractions intruded as discordant stocks.

By 1950 Browne (Browne in David, 1950) also had realised that the Kosciusko Batholith was composite but, after having noted that both 'biotitic' and 'hornblendic' granitoids were present, he proceeded to use a textural criterion in proposing a three-fold subdivision of the Lachlan Zone plutonics. Each type he considered to be the product of a particular orogenic episode, which occurred in the Ordovician for the 'granite gneisses', in the Silurian for the granitoids which possessed a 'gneissic foliation partly but not entirely cataclastic' and in the Devonian or later for those which were 'massive and unstressed'.

The same classification and relative ages were adopted by Joplin (1962) but she considered that the different granitoid types were produced as a function of the phase, or time, of the development of the granitic magma within a single orogenic epoch. She distinguished the granitoid types geochemically and pointed out the high CaO and Na$_2$O contents and the low K$_2$O/Na$_2$O of the 'massive' granitoids relative to the 'foliated' and 'gneissic' types.

Joplin's work was followed up by the much more comprehensive study of Kolbe and Taylor (1966). Their geochemical examination of the granitoids in the Snowy Mountains area was undertaken expressly 'to examine the hypothesis of Joplin (1962)' and 'to provide data that may give an answer to some of the problems of granite genesis in general' (pp. 1-2). In retrospect, their success in both regards was limited, principally due to the difficulties in obtaining sufficient geological control on the sampling. They concluded that, despite the fact that 'gneissic granites' and 'leucogranites' could be distinguished from one another and from the
'granodiorites and adamellites', no chemical subdivision of the lattermost which coincided with the petrological classification of 'massive' and 'foliated' could be made.

In the same year, Chappell (1966) completed an investigation of the petrogenesis of the granitoids at Moonbi, part of the New England Batholith, 700 kilometres to the north. His conclusions shaped the philosophy behind most of the later chemical studies undertaken in the southern batholiths. Chappell advanced an argument based upon the chemical analysis of nearly fifty granitoid and xenolith samples, which established the existence of a direct relationship between the composition of the granitoid and the composition of its source material. In essence, he demonstrated that the greater proportion of the xenoliths found in the Moonbi granitoids were cognate and employed that, and other facts, in support of the hypothesis that the chemical heterogeneity observed within any one pluton, and the differences between members of a suite of related plutons, were a function of the mixing of felsic magma with different proportions of solid mafic material at the source. (In the light of Chappell's work, the term 'xenolith' was a misnomer but it is so entrenched in granitoid nomenclature that it will continue to be used here to describe inclusions in granitic magmas. The distinction between inclusions which are cognate and those which are xenoliths in the strict sense of the word will be made by using the terminology used by Tyrrell (1926) and referring to the latter as 'accidental xenoliths').

Working from his hypothesis, Chappell was able to estimate the composition of the granitoids' source material. For the majority of the granitoids it proved to be shoshonitic but in one case he considered that partial melting of a mixture of basic material and sialic crust may have taken place.

Chappell's concept was soon afterwards applied to the southern batholiths, where the question of differences in granitoid source materials had remained unanswered for a considerable time. If granitoids which
contained mafic igneous xenoliths were derived from an igneous source, it was likely that granitoids which were rich in metasedimentary xenoliths could be shown to originate from a metasedimentary source.

Joyce (1970, published Joyce, 1973) analysed granitoids and xenoliths from the Murrumbidgee Batholith, which previous petrographic work by Snelling (1960) had shown to be a composite of several intrusives, most of which contained xenoliths that were apparently of sedimentary origin. Joyce retained Snelling's classification of the granitoids as 'contaminated' and 'uncontaminated' (in essence, rich or poor in metasedimentary xenoliths respectively) but noted the need for a classification with less genetic implications. Granitoids in both these categories had been classified earlier by Joplin (1962) as 'foliated'.

The conclusions from Joyce's work were that an origin by partial melting of psammopelitic source rocks, followed by varying degrees of reaction with solid relict material (represented by the xenoliths) at, or near the source, was the explanation for the observed compositional range in the great majority of the rocks in the Murrumbidgee Batholith. The one 'massive' granitoid in the Batholith he considered to be analogous to the Moonbi granitoids in being of igneous derivation.

While Joyce worked on the Murrumbidgee Batholith, Chappell, White and their students were concluding the detailed geological mapping of the Berridale Batholith (Lambert, 1963; Lambert & White, 1965; Black, 1965 and Patterson, 1968) and geochemical work on the newly defined intrusives was begun. At the same time, other students worked on the adjacent Kosciusko Batholith (Hine, 1971 and Williams, 1973). From these studies a new classification scheme arose, but one which had a firm geochemical foundation, the concept of I- and S-type granitoids.

1.4.2 I- and S-type granitoids

The geochemical studies of the Kosciusko and Berridale Batholiths demonstrated that the 'massive' - 'foliated' classification of Dallwitz
(Ball et al., 1948) was reflected in geochemical differences between the granitoids and that there were many more differences than those in major element composition which Joplin (1962) had proposed. It was found also, however, that there was not an invariant correlation between massive and foliated textures and the mineralogical criteria on which Dallwitz's classification rested (Williams, 1973). A nomenclature divorced from textural considerations was therefore coined. The 'massive' granitoids were termed 'I-type' and the 'foliated' granitoids 'S-type' (Chappell & White, 1974), these names recognising differences in the nature of the inferred source material, Igneous and Sedimentary rocks respectively.

The classification of a particular granitoid as I- or S-type rested ultimately upon its chemical composition. Nevertheless, the chemical differences between the two classes correlated with differences in the mineralogy which, in all but a few examples of very felsic rocks, could be used in making a preliminary assignment. A recent summary of the rapidly-growing list of petrographic characteristics of the two types has been made by Hine et al. (1976), who discussed the application of the classification to the granitoids of the eastern side of the Kosciusko Batholith. The most important of the characteristics were as follows.

1) **I-type granitoids and their xenoliths**

   Hornblende and biotite are the dominant ferromagnesian minerals.
   
   Biotite is pleochroic straw-yellow to dark sepia-brown.
   
   Plagioclase is abundant, commonly euhedral and oscillatory zoned from a rounded core about An$_{80}$ to a rim An$_{40}$ to An$_{25}$.
   
   Quartz is grey and shows only weak undulose extinction.
   
   Magnetite is a common accessory phase (as elsewhere is sphene (Chappell & White, 1974)).

2) **S-type granitoids and their xenoliths**

   Hornblende is absent but biotite may be very abundant.
   
   Biotite is pleochroic straw-yellow to foxy red-brown.
Cordierite is common in the more mafic granitoids.

Alumino-silicates and garnet may be present (Chappell & White, 1974).

Muscovite may occur as a primary phase.

Plagioclase forms rectangular zoned crystals with cores about An$_{55}$ and mid-oligoclase rims.

Quartz is commonly lilac-blue and always has strongly undulose extinction.

These mineralogical characteristics are only a reflection of more fundamental differences in the granitoids' chemistries and modes of origin. The major chemical differences and the manner in which they arise may be summarised as follows (Chappell & White, 1974; White et al., 1977; White & Chappell, 1977; Hine et al., 1976; Griffin et al., 1976).

1) S-types are depleted relative to I-types in those elements which are lost from sediments during the sedimentary cycle (for example Na, Ca and Sr) and enriched in those which are gained (for example K, which is adsorbed onto clays).

2) As a consequence of the above, S-types are peraluminous in the sense of Shand (1950) namely, $\text{Al}/(\text{Na}+\text{K}+\text{Ca}/2)$ is greater than 1.1, but I-types are not.

3) I-types display a broad spectrum of compositions from felsic to mafic and show interelement variations which commonly are linear. S-types have a more restricted range of silica contents and interelement variations are more irregular.

4) S-types generally are the more reduced types, presumably because of carbon or sulphur in the original sediments, and therefore have low $\text{Fe}_2\text{O}_3/\text{FeO}$.

5) S-types, relative to I-types, are rich in Fe, Mg and Ti for a given silica content, also in other ferromagnesian-type elements (for example Co, V, Se, Ni and Cr) and in elements which behave like K (for example Rb, Th and U).
1.4.3 **Isotopic studies**

1.4.3.1 **Radiometric isotopes**

The principal objective of isotopic studies of the southeastern granitoids has been, until recently, the measurement of geological time. Before the advent of radiometric dating, stratigraphic relationships were the sole means of determining the age of the intrusives and conclusions reached in one area were extended to others by the correlation of lithologies (Browne, 1929, 1931; in David, 1950). Radiometric age determinations have demonstrated the difficulties with such an approach (Evernden & Richards, 1962).

Evernden and Richards (1962) undertook a reconnaissance survey of the igneous rocks in southeastern Australia, measuring ages by the potassium-argon technique. Theirs was pioneering work in Australia and, despite the very low density of sampling and common lack of geological control, established the broad pattern of igneous activity over the whole area, a pattern which has remained fundamentally unaltered to the present. The most important of their conclusions were as follows:

1) The 'age-type' nomenclature which had been used by Browne (1929), for example, was not borne out by the new data and its abandonment was recommended.

2) The granitic rocks fell naturally into two groups, the first, Siluro-Devonian granitoids which formed intrusive masses with 'a more or less meridional outline' and the second, 'irregular smaller Carboniferous masses of varying trends' (p. 18).

3) The measured ages indicated an 'eastward migration of the axis of intrusion' which was 'astonishingly clear' (p. 20). The lines of intrusion were parallel to the gross structural features of the Tasman Zone and 'the axes fellow[ed] in perfect order' from Siluro-Devonian through to the early Cretaceous.

4) The correlation of plutonism with major orogenies was not possible and the correlation of particular rock types with particular
orogenies was 'forced and inaccurate' (p. 20).

Geochronological work succeeding that of Evernden and Richards has been directed towards the solution of more local problems. Commonly, it has involved not only an attempt to date the igneous activity but also the employment of isotopes as tracers in deducing something of the pre-intrusive history of the magmas. The decay scheme used most widely in this regard has been rubidium-strontium.

The use of strontium as an isotopic tracer (as distinct from its use in measuring geological ages) stems from the work of Gast (1960). He reasoned from the relatively low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios measured in a selection of basaltic rocks which were considered to be mantle derived, that the rubidium to strontium ratio in the upper mantle was lower than that in the crust. This being so, igneous rocks originating in the crust would be expected to have characteristically higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than those originating in the upper mantle.

This prediction was put to the test by Hurley et al. (1962) who examined the isotopic composition of strontium in felsic plutonics and volcanics, schists and gneisses from several countries. Considering the granitic rocks from North America as representative of a 'typical continent', Hurley et al. showed that very few had initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios markedly higher than the average value of 0.708 (which was also the average value for oceanic and continental basalts reported by Faure & Hurley, 1963) and none was within the range 0.720 to 0.730 which they expected for reworked ancient crust. As a consequence, it was concluded that the hypothesis of reworking of ancient crust to form these rocks had to be rejected.

This opinion was not shared by all workers, however, for cases were known outside North America in which the intimate relationship between granitic rocks and high-grade metamorphics suggested strongly the formation of the granitic rocks by anatexis. One such example, the Cooma Granite, was the subject of a study by Pidgeon and Compston (1965).

The Cooma Granite forms a small, isolated pluton located between
the Murrumbidgee and Berridale Batholiths. It is unique in that area in being surrounded by a regional metamorphic aureole with an outcrop area an order of magnitude larger than that of the granitoid itself. The host rocks are metasediments and metamorphism grades from an outermost chlorite zone through to an inner migmatite zone, in the centre of which is the Granite. It is likely that the granitic magma has in fact been mobilised (Joplin, 1962) but its derivation from a metasedimentary source is indisputable (Chappell & White, 1976).

Pidgeon and Compston (1965) analysed both minerals, and total rocks sampled from the Granite and the metasediments of the aureole, for rubidium-strontium. All samples from the Granite were found to be colinear when plotted on an isochron diagram and gave an apparent age of 415 ± 12 m.y. and an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7179 ± 0.0005 which they considered to be 'usually high'. Measurements on the samples from the surrounding high-grade metasediments gave apparent ages identical within error to that of the Granite and similarly high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (which nevertheless varied significantly from one zone to another). In contrast, the greenschist-facies rocks had a significantly greater isotopic age and lower initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio.

On the evidence of a difference in the apparent ages between the high- and low-grade metamorphics, Pidgeon and Compston rejected the hypothesis that the whole area had suffered a post-emplacement, high-grade metamorphic event and concluded that the Granite isochron was giving a true indication of the rock's age of formation and initial strontium isotopic composition. That being so, the high initial content of radiogenic strontium was considered either to be inherited from the Granite's source material or to be the result of the assimilation of crustal rocks during emplacement. The analytical data was thus compatible with the hypothesis of anatexis.

Although the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the Cooma Granite was high with respect to values observed in granitoids elsewhere, it was still below
those predicted by Hurley et al. (1962); the search for even higher levels of initial radiogenic strontium in granitoids continued. They were found in the Heemskirk Granite, western Tasmania (Brooks and Compston, 1965).

The Heemskirk Granite is unusual in comparison with other granitoids of the Tasman Zone in being in contact largely with Precambrian metasediments. The margin of the Granite is faulted in places but elsewhere, a contact aureole is present and the body is demonstrably intrusive. The Granite is not of uniform composition and may be subdivided into at least three phases on the basis of the colour and form of the constituent orthoclase. The internal contacts between the phases are zones of irregular, diffuse mixing.

Brooks and Compston's (1965) rubidium-strontium analyses of total-rock samples from the Granite disclosed significant differences in the composition of the initial strontium both between and within the phases, and values for the initial $^{87}\text{Sr}/^{86}\text{Sr}$ so high (0.716 to 0.741) that the post-crystallisation migration of radiogenic strontium had to be considered as a possibility. This they did, and concluded that there had been neither redistribution of radiogenic strontium during a late metamorphic event nor addition of radiogenic strontium to the Granite subsequent to its emplacement. They favoured instead either the generation of the magma from crustal rock which was already enriched in radiogenic strontium or the assimilation of strontium from outside sources by the Granite prior to its crystallisation; the data did not permit a distinction between those alternatives. The use of initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the study of magma genesis has now become worldwide. The review compiled by Faure and Powell (1972) details further examples.

Strontium isotopic studies have been an integral part of the geochemical work on the southeastern Australian batholiths but they have tended to post-date the major and trace element work and are subsequently still very much 'in progress'. One batholith for which the data has been fully assessed is the Murrumbidgee Batholith (Roddick, 1974). Aspects of
Roddick (1974) analysed total-rock and mineral samples from several plutons within the Murrumbidgee Batholith and therefore was able both to compare the mineral and total-rock ages and to detect differences in the composition of the initial strontium within and between intrusives. The isotopic data could not be interpreted simply; in addition to scatter on an isochron plot of some of the analyses which was far beyond that accounted for by experimental error, there were several cases in which the total-rock samples defined isochrons registering significantly greater apparent ages than did the minerals. Roddick's explanation of such analyses had wideranging implications for future total-rock work on similar material.

With one exception, the northern phase of the Tharwa Adamellite, the mean initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the plutons were within the range 0.709 to 0.715. This was not a great deal lower than the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the Cooma Granite (0.7179), and was considered to be compatible with the postulate of Joyce (1970) that the granitoid source rocks had been metasediments. After considering the work of Compston and Pidgeon (1962) and Bofinger et al. (1970) in which rubidium-strontium ages were measured on sedimentary rocks, Roddick suggested that the metasedimentary source of the granitoids may have preserved a record either of its deposition, or latest diagenesis, in its distribution of rubidium and radiogenic strontium, and that that record may have been 'inherited', in whole or in part, by the derivative granitoid. The apparent age of the 'inherited isochron' in such a case would have time significance only if the rubidium to strontium ratio in sample-sized volumes had not been altered during granitoid formation and this was considered to be most unlikely.

Roddick's work demonstrated that the processes by which the Murrumbidgee Batholith granitoid magmas were generated did not always result in the equilibration of the strontium in the melt with that in the solid residue (or 'restite' in the terminology of White & Chappell, 1977). In
such cases, the time of emplacement would be more likely to be recorded, not by the total-rock, whose apparent age would be a maximum estimate, but by the minerals.

The results from the Murrumbidgee Batholith are not an isolated example of 'inherited' rubidium-strontium total-rock ages. Several groups reporting rubidium-strontium analyses of lunar basalts from the Ocean of Storms (for example, Compston et al., 1971) have observed a strong alignment of data along a 4.5 b.y. isochron, in complete contrast to the mineral ages of individual samples which were in the range 2.9 to 3.6 b.y. Hypotheses that can explain this observation are first, variable contamination of the basalt melts with 4.5 b.y. old material which was comparatively rich in radiogenic strontium, second, total extraction of rubidium and strontium from the source of the basalts during melting, with the implication that the rubidium-strontium ratio in the source remained heterogeneous from 4.5 b.y. ago, and third, disequilibrium melting.

Graham and Ringwood (1971), who were concerned with explaining europium anomalies in the same rocks, noticed that the rare earths and other incompatible elements (Ba, U, Zr and Th for example) all had displayed generally similar crystal-liquid partition coefficients during magma genesis. This behaviour could not easily be explained in terms of equilibrium crystal-liquid partition coefficients involving the major phases in the source region so they suggested that the incompatible elements in fact dominantly were in low melting point accessory minerals. If that were so, with the batch melting of these minerals to form the first liquids, the incompatible elements would be transferred to the melt quantitatively. Larger degrees of partial melting would simply cause dilution, on the condition that the melt moved away from the parent crystalline phases as it formed and that the incompatible elements remained out of equilibrium.

One of the elements to concentrate in the accessory phases would be rubidium (Compston in Graham and Ringwood, 1971; Compston, 1974) and they would as a consequence accumulate radiogenic strontium. The first
formed liquid would thus contain virtually all the rubidium and radiogenic strontium. Common strontium, being dispersed through the major phases, would dilute the radiogenic strontium as the degree of partial melting increased, so generating an array of total-rock compositions defining an isochron and hence the pre-melting age. This model may not be applied directly in the explanation of results of analyses of felsic melts, like those in the Murrumbidgee Batholith, but the example does demonstrate that melt and solid source residues do not always attain mutual isotopic equilibrium.

1.4.3.2 Stable isotopes

Stable isotopes in the southern batholiths have received less attention than radiogenic isotopes. The single study published is that by O'Neil and Chappell (1977) of the isotopic composition of oxygen and hydrogen in the Berridale Batholith.

O'Neil and Chappell demonstrated the close correspondence between the classification of the granitoids as I- or S-type and their oxygen and hydrogen isotopic compositions. The oxygen isotopic composition within individual I-types was far more homogeneous than that in S-types but the total range in, $\delta^{18}O$ values for whole-rock samples, all plutons considered, was much greater (7.9 to 9.4 per mil in I-types and 9.9 to 10.5 per mil in S-types). The ranges of $\delta^{18}O$ values in S- and I-types did not overlap. Hydrogen isotopic compositions showed a similar pattern to those of oxygen, but the differences between the granitoid types were not as sharply defined.

Rather than invoking exchange processes between granitoid and enclosing rocks, O'Neil and Chappell explained their data by differences in the granitoids' sources. They regarded the observed isotopic composition as 'a remnant of the original material which was partially melted to form the granitoid'. The range in measured oxygen and hydrogen isotopic compositions was explained by different degrees of differentiation and crustal contamination of the magmas.
1.5 GRANITOID GENESIS

Chappell's (1966) concept that granitoid magmas were a mixture of chemically related melt and solid source residues was expanded into a single theory of granitoid genesis by White and Chappell (1977). Their proposal was that a granitic magma is a mixture of the products of ultrametamorphism: a minimum or non-minimum melt and residuum or 'restite'. Any granitoid, with the exception of those which had been totally molten (the Tuolomne Suite, California, for example) was considered to contain relict solid material from its source, either as large inclusions (cognate xenoliths) or disseminated as individual crystals through the rock.

White and Chappell regarded the linear Harker variation diagrams produced by analyses of suites of related samples (as reported by Hine et al., 1976 and Griffin et al., 1976, for instance) as mixing (more strictly 'unmixing') lines generated by magmas which had been freed of restite to varying degrees. The composition of the source was between that of the two end-members, the solid restite on the one hand and the melt on the other.

By such a model, the apparent anomalies in rubidium-strontium total-rock ages reported by Roddick (1974), Roddick and Compston (1977) and Compston and Chappell (1976), for instance, might possibly be explained. If the melt and restite did not attain strontium isotopic equilibrium and their rubidium to strontium ratios were different, the process of 'unmixing' would generate an 'isochron' in the magma before it was emplaced. Compston and Chappell (1976) concluded that for the general case, the model predicted initial isochrons with positive slopes, because the felsic melt would be richer in alkalis, and therefore in radiogenic strontium, than the solid residue. Although this put the usefulness of the rubidium-strontium total-rock method in measuring granitoid emplacement ages further in doubt, a study of source rock ages became possible.
Compston and Chappell (1976) combined an estimate of the rubidium to strontium ratio in the source, calculated from the geochemistry, with the measured initial $^{87}\text{Sr}/^{86}\text{Sr}$ in several related plutons to reconstruct the rubidium-strontium isochron parameters of the source region. For the Berridale Batholith, they obtained an approximate estimate of 625 m.y. for the source age of one suite of I-type plutons.

1.6 THE OBJECTIVES OF THE PRESENT STUDY

In early 1974, when the present project was begun, Roddick (1974) had established the age anomalies in the Murrumbidgee Batholith, and the combined work of Compston, Chappell and Shirahase (unpublished) had provided convincing evidence for at least one case of an even larger anomaly in the Berridale Batholith. It was proposed that a study of the Berridale Batholith be undertaken to determine the granitoids' emplacement ages and, if possible, the ages of their source rocks.

The problem was approached from three directions. First, some of the earlier rubidium-strontium total-rock analyses were duplicated and additional samples were collected from problematical plutons. The results of that work will not be discussed in detail in this thesis for it did not constitute a self-contained study, and all the total-rock data are soon to be reported elsewhere (Compston et al., 1977a). Preliminary total-rock ages calculated from the data of Compston et al. (1977a) are included however.

Secondly, rubidium-strontium mica ages were measured over the entire batholith. The results have been reported by Williams et al. (1976). The study of the mica ages was co-ordinated with the work of another postgraduate student, N.W. Tetley, who used many of the same samples for age determinations by potassium-argon and $^{40}\text{Ar}^{39}\text{Ar}$. Unless otherwise acknowledged, the potassium-argon ages discussed in this thesis are included with his kind permission. Thirdly, and most importantly, selected granitoids were dated by zircon and monazite uranium-lead.
The uranium-lead decay scheme was one of the first to be used in the measurement of geological time (Rutherford, 1929; Nier, 1939 and Nier et al., 1941). The scheme has the unique advantage that it is paired; two of the isotopes of uranium (\(^{235}\)U and \(^{238}\)U) decay by different paths and at different rates to two of the isotopes of lead (\(^{207}\)Pb and \(^{206}\)Pb respectively). An age can be calculated from the lead composition (\(^{207}\)Pb/\(^{206}\)Pb) as well as from either of the daughter-parent ratios (\(^{207}\)Pb/\(^{235}\)U and \(^{206}\)Pb/\(^{238}\)U).

Early workers with uranium-lead found that the ages calculated from these three isotopic ratios commonly disagreed. By the time that Kulp et al. (1954) published their review of the lead method of age determination, it had become apparent that in most cases the nature of the disagreement was the same; the \(^{207}\)Pb/\(^{206}\)Pb age was considerably higher than the \(^{207}\)Pb/\(^{235}\)U age, which was in turn slightly higher than the \(^{206}\)Pb/\(^{238}\)U age. Soon afterwards, Ahrens (1955a) demonstrated that the disagreement was systematic.

Ahrens (1955a) examined uranium-lead data from monazites from the Rhodesia-Tanganyika Shield (Holmes, 1954), and by plotting the ages calculated from \(^{206}\)Pb/\(^{238}\)U, \(^{207}\)Pb/\(^{235}\)U, \(^{207}\)Pb/\(^{206}\)Pb and \(^{208}\)Pb/\(^{232}\)Th against the \(^{207}\)Pb/\(^{235}\)U age (as a reference), he obtained a systematic relationship which he termed the 'Rhodesia age pattern'. The ages calculated by the four methods plotted on four curves which converged at a point, the 'convergent age'. Ahrens interpreted the pattern as being generated by the physical loss of lead and considered the convergent age to be a better estimate of the true age than that calculated from any one of the isotopic ratios.

In a subsequent paper Ahrens (1955b) used a semi-log scale for his plot and showed that the isotopic ages then defined straight lines. The line of 'age equality' however, became a curve. He showed also, that straight lines could be obtained on a linear scale if the measured isotopic ratios, rather than the calculated ages, were plotted.
Wetherill (1956a, b) took Ahrens' interpretation one step further. He noted that the lines defined by the data on a $^{207}\text{Pb}/^{235}\text{U}$ vs. $^{206}\text{Pb}/^{238}\text{U}$ plot intersected the curve of age equality (which he called the 'concordia') at two places. He proposed that the upper intersection indicated the true age of the samples and that the lower intersection indicated the time at which they had suffered an episode of lead loss.

Following this work, lead loss became a central issue in the interpretation of uranium-lead data, which even now has not been finally resolved. Two fundamentally different explanations of lead loss were proposed. The first was that lead loss from a mineral is continuous from the time it forms (Nicolaysen, 1957; Tilton, 1960; Wasserburg, 1963; Wetherill, 1963 and Ulrych, 1963, for example). The second was that lead loss occurs only in response to an episodic change in physical or chemical conditions (Aldrich & Wetherill, 1958; Silver & Deutsch, 1961, 1963; Wasserburg et al., 1962 and Catanzaro & Kulp, 1964, for example). A correct explanation of lead loss was vital to the interpretation of the lower intersection of the line defined by the data (the 'discordia') and the concordia. By the episodic model, this represented the age of an episode but by the continuous diffusion model, it was an apparent age only, for the discordia was not a straight line but a curve.

The approach adopted in the analysis of zircons changed with the work of Silver and Deutsch (1961, 1963). They demonstrated that the zircon in one rock had a range of chemical and isotopic compositions and that a discordia could be defined by a single sample. They did this by utilising a correlation between the degree of discordance of a particular zircon fraction and its total radioactivity. In this way a discordia could be generated while cogenetic of the samples was guaranteed.

In 1968, Grauert and Arnold (1968) analysed zircons from paragneisses in the Swiss Alps and found a pattern of discordance which had not been observed previously. Zircons from nearby igneous rocks and orthogneisses gave ages in the interval 400 to 500 m.y. but the paragneiss
zircons, using an episodic lead loss model, indicated only a disturbance at that time. The primary age was 1500 m.y. Grauert and Arnold concluded that 1500 m.y. was in fact the age of a detrital zircon component.

This phenomenon of inheritance has since been observed in many studies of zircons from granitoids and gneisses. Most of these are studies done in western Europe (for example, Michot & Deutsch, 1970; Pidgeon et al., 1970; Köppel & Grünenfelder, 1971; Gulson & Krogh, 1973) but inheritance has also been reported from North America (Grauert & Hofmann, 1973).

By analysing zircons from the Berridale Batholith, it was hoped to take advantage of inheritance. The rubidium-strontium work of Compston and Chappell (1976) had shown that the source rocks for the granitoids were significantly older than the plutons themselves. If the granitoid zircon populations were a mixture of zircons from the source and zircons crystallised at the time of emplacement, both source age and emplacement age might be determined. There had in fact been one zircon study done in the Snowy Mountains region (Richards et al., 1966) and a single anomalously high zircon lead-alpha age had been found.
CHAPTER 2

SELECTION OF SAMPLES

Because it was not possible to sample every granitoid in the Berridale Batholith for zircon, it was necessary to select those granitoids which were potentially of the greatest geochronological interest. The selection was made on the basis of what was already known about the granitoids' field relationships, geochemistry and geochronology.

From intrusive relationships, it was possible to determine the relative emplacement ages of many of the plutons. These are illustrated diagrammatically in Fig. 2.1. The oldest pluton was thought to be the Finister, an I-type granitoid which, with the Merumbago, was considered to be geochemically distinct from all the other I-types in the Batholith (White et al., 1978).

After the Finister, a group of plutons consisting of the Numbla Vale, the Dalgety, the Cootralantra (three S-types) and the Buckleys Lake (the largest of the I-types) was emplaced. Within this group, an intrusive sequence has not been established. The Cootralantra is intruded by the Buckleys Lake, which in turn is intruded by the Dalgety. However, at two different localities on the boundary between the Cootralantra and the Dalgety, conflicting relative ages have been determined (White et al., 1977). This does not imply that either determination is a misinterpretation of the field relationships; the Cootralantra intrusive is a complex of many plutons and may have been emplaced over an extended period of time which spanned the time of emplacement of the Dalgety.

The Cootralantra and the Buckleys Lake are intruded by three younger plutons, each of which is an I-type; the Tara, the Maffra and the Wullwye. The three are not in mutual contact so it has not been possible to use field relationships to determine their relative ages.
Fig. 2.1 Intrusion relationships in the Berridale Batholith. The oldest pluton is the Finister. It was followed by the Numbla Vale, the Dalgety, the Buckleys Lake and Cootralantra. Field evidence on the relative ages of the Dalgety and the Cootralantra is conflicting (see text). The youngest plutons are the Tara, the Maffra and the Wullwye. The Delegate is not in contact with any other granitoid.
INTRUSION RELATIONSHIPS

FIGURE 2.1
When the sampling of the Batholith for zircon was planned, granitoids from each of these age groups were selected. In addition, within each group, those granitoids which represented the extremes of the observed chemical range were chosen. The plutons which were sampled and the reasons for their selection were as follows:

**The Finister Granodiorite**

The Finister was considered to be the oldest pluton in the Berridale Batholith. First, it appeared to be an integral part of a narrow, otherwise metasedimentary, screen which was part of the rest of the Batholith's country rock; secondly, it was intruded by the Buckleys Lake and the Numbla Vale which themselves were considered to be relatively old, and thirdly, its rubidium-strontium total-rock age exceeded that measured on any other granitoid.

The Finister was of particular interest from a geochemical point of view. Its chemical composition was so distinctive that it had been proposed that the magma which formed the Finister and the Merumbago had come from a different source than had the magmas which formed the other I-types (White et al., 1978). It was important to use zircon analyses in an attempt to establish whether this chemical difference corresponded to a difference in the sources' ages.

A further reason for determining the Finister's zircon age was that the results of the geochronology which had been done on the pluton were ambiguous. The rubidium-strontium total-rock age was so much greater than the total-rock age of any other pluton that its value as a measure of the age of emplacement was suspected. This suspicion was enhanced by an apparent contradiction between the total-rock age and the postulated age of the sediment which the pluton intruded. Age determinations on the Finister by techniques such as potassium-argon or rubidium-strontium on minerals were thought to be unlikely to give the pluton's age of emplacement because of the proximity of large, later plutons. Zircon
uranium-lead was thought to be one technique which might escape resetting and record an emplacement age.

**The Delegate Adamellite**

A second pluton which had an old rubidium-strontium total-rock age was the Delegate, but in contrast to the Finister, its measured age was compatible with the stratigraphic constraints. Because the Delegate is isolated from the rest of the granitoids in the Batholith (Fig. 2.1) it was not possible to determine its relative emplacement age from intrusive relationships. The only way that the total-rock age could be confirmed as being the age of emplacement was to duplicate it by another isotopic technique. Zircon uranium-lead was thought to be the most appropriate. If the Delegate really was as old as its total-rock age suggested, the majority of the rest of the Batholith postdated it, so a method resistant to resetting was required. If, on the other hand, the total-rock age was partially inherited from an older source rock, zircon uranium-lead might be a means of determining how old that rock was.

**The Tara Granodiorite and Maffra Adamellite**

The two other I-types which were selected were the Tara and the Maffra. Those particular plutons were chosen in preference to any of the six other I-types in the Batholith for three main reasons:

1) Each is in intrusive contact with one of the major plutons in the Batholith (the Cootralantra and the Buckleys Lake respectively) so their relative emplacement ages were known.

2) Both plutons were known to be geologically relatively young and their rubidium-strontium total-rock ages were also.

3) The Tara and the Maffra represented the two extremes of the range of chemical compositions observed in the three youngest I-types which were discussed above.

The Maffra is the most felsic I-type granitoid in the Berridale
Batholith, with the possible exception of numerous small bodies of aplite. It is virtually free of mafic xenoliths and almost of minimum melt composition (Chappell & White, 1976). By the 'unmixing' model of granitoid genesis, the Maffra was considered substantially to have freed itself of solid residual source material and it was therefore expected to contain little or no zircon inherited from that material. The Maffra was chosen as the pluton in which it was most likely that the zircon would record only the emplacement age.

In contrast to the Maffra, the Tara is a mafic granitoid (transitional between granodiorite and tonalite) in which mafic xenoliths are very abundant. On both chemical and mineralogical grounds, the amount of residual source material in the granitoid magma was thought to have been very large, but in addition, the total-rock rubidium-strontium analyses indicated that the granitoid had a very uniform initial strontium isotopic composition. The Tara magma was therefore considered to have sampled effectively a relatively homogeneous source. Since that sampling was expected to include zircons from the source, the Tara appeared to provide an opportunity to date zircons from igneous source material similar to that of most of the other I-types.

S-type granitoids also were sampled. Similarly to the sampling of the I-types, those plutons were selected which would be of greatest use in solving first order geochronological problems, but at the same time, an attempt was made to sample as wide a compositional range as possible. The zircon results from the S-types were expected to be more complex than those from the I-types. The S-types' source was considered to be metasedimentary and the zircons in it could not be assumed all to have the same provenance.

The Cootralantra Granodiorite

The most abundant and most mafic S-type granitoid in the Berridale Batholith is the Cootralantra. It is characterised by being
rich in xenoliths, the great majority of which preserve readily recognizeable features of the metasediment from which they are derived. As in the case of the Tara, the Cootralantra was considered to be mafic due to the high proportion of residual source material which had been retained by the magma. This reasoning was supported by the rubidium-strontium total-rock analyses from the pluton. They indicated a heterogeneity in initial strontium isotopic composition even larger than that reported in S-type granitoids of the Murrumbidgee Batholith by Roddick (1974). It was therefore anticipated that the Cootralantra would contain a large amount of detrital zircon inherited from its source and that the age of the source rock's provenance might be determined. Moreover, if the remainder of the zircon in the Cootralantra crystallised on emplacement, the age of that event might also be obtained.

**The Bullenbalong Granodiorite**

A second granitoid sampled principally for the information that it might provide about the age of the S-types' source was the Bullenbalong. In appearance and major element chemistry, the Bullenbalong closely resembles the Cootralantra, however it is part of the Kosciusko, not the Berridale Batholith. As with the Cootralantra, the Bullenbalong was thought to contain a large amount of solid residue from its source and part of that residue was expected to be detrital zircon. The Bullenbalong made it possible therefore, to compare zircons derived from different parts of the S-type source but under similar magmatic conditions.

**The Dalgety Granodiorite**

Although it is an S-type granitoid and only slightly more felsic than the Cootralantra, the Dalgety has a texture which indicates that it evolved in a very different way. In contrast to the Cootralantra's irregular texture and aggregated minerals, the Dalgety is coarse-grained, relatively even-textured and, in most places, is composed of well-formed
crystals. In addition, the Dalgety contains very few xenoliths and those that are present do not have the appearance of pieces of metasediment. This texture was interpreted as indicating that the Dalgety formed by a relatively high degree of partial melting and that the material which in the Cootralantra had remained as solid residue, in the Dalgety had been incorporated into the melt phase.

The interest in studying zircon from the Dalgety, apart from measuring its emplacement age, was to determine whether a higher degree of partial melting had an effect on the amount of zircon which was inherited from the source. In addition, it had been suggested (White et al., 1978) that the Dalgety was chemically sufficiently different from the Cootralantra (and the other S-types) for its derivation from a different source to be very likely.

**The Numbla Vale Adamellite**

The investigation of source-derived zircons in a melt (in comparison to a melt-solid mixture) was continued through a study of the Numbla Vale. The Numbla Vale, like the Dalgety, contains very few mafic xenoliths but in contrast to the latter, its chemical composition is very close to that of a minimum melt (Chappell & White, 1976). In this respect, the Numbla Vale was considered to be the S-type equivalent of the I-type Maffra. The Numbla Vale was thought to have freed itself of virtually all solid source residue and it therefore was expected to contain considerably less zircon inherited from its source than was contained in the Cootralantra or the Bullenbalong. If most of the zircon was crystallised at emplacement, the Numbla Vale zircon age would be an important upper bound on the emplacement age of the Finister.

**The Bog Hole Granite**

The Bog Hole, like the Bullenbalong, is a granitoid from the Kosciusko Batholith. It was sampled because it was thought to have
formed from a magma which contained even less solid residual source material than the magma which formed the Numbla Vale. Any zircon in the Bog Hole which was not crystallised from the melt would have had to have come from the granitoid source in the melt phase. A study of the zircon was expected to show to what extent this process occurred.

Xenoliths

In addition to granitoid samples, two samples of xenoliths were collected from the Berridale Batholith; one was a piece of cordierite gneiss (which was clearly of metasedimentary origin) from the Cootralantra, and the other was an even-textured, mafic xenolith from the Tara. The xenoliths in both cases were considered to be pieces of modified solid residuum from the granitoid's source. It was anticipated that they would be strongly enriched in zircon from the source and therefore of value, not only in dating the source, but also in identifying the source component in the analyses of zircons from the granitoid itself.

SUMMARY

Zircon samples from the Berridale Batholith were selected with the aim of determining both emplacement ages and the ages of the granitoids' source materials. The granitoids chosen were both I- and S-types and had a range of chemical compositions and geological ages, the intention being to study granitoids which had been derived from different source materials and by different degrees of partial melting, and which contained different amounts of solid source residuum.
CHAPTER 3

THE MAFFRA ADAMELLITE

3.1 THE GRANITOID

The Maffra Adamellite has been described by White et al., (1978). In hand specimen, it is a fine-grained, even-textured rock. Its content of ferromagnesian minerals is very low (biotite alone is present and accounts for only 3.1 percent of the mode) so the dominant coloration is the pink of the potassium feldspar. Chappell and White (1976) have observed that in the Berridale region, pink feldspar is found only in I-type granitoids. In very fresh specimens, the plagioclase has a distinctive blue-green tint.

The Maffra's chemical composition is very close to that of a minimum melt. This, in combination with a virtual absence of mafic xenoliths, suggests that the magma either left its source as an almost pure minimum melt or that it freed itself of solid mafic residuum during its ascent. By either interpretation, the amount of zircon introduced into the granitoid magma by the assimilation of solid residual source material can be only very small.

3.2 THE ZIRCON

The zircon sample of the Maffra (75-117) was collected from the same locality as the geochemical sample BB3l. In thin section, the zircons are found to occur as discrete grains included in the major mineral phases. Their distribution is not uniform, however (Fig. 3.1 ). Few zircons are found included in quartz, and those that are present are very small. Zircons are present in the highest concentration in biotite, but also occur in potassium feldspar and plagioclase. No zircon has been seen in plagioclase cores. Independently of where they occur, the majority of the zircon grains are closely associated with opaques and/or euhedral apatite. Many of the zircons appear clouded and corroded.
Fig. 3.1 The distribution of zircon in the minerals of the Maffra Adamellite, as observed in thin section. Each grain is drawn to scale. Blackened grains are those which were closely associated with grains of apatite and/or opaque minerals. The dashed area in the plagioclase represents the cores of plagioclase grains. The area allocated to each of the major minerals is in proportion to that mineral's modal abundance.
FIGURE 3.1

MAFFRA ADAMELLITE

100μ

K-FELDSPAR

PLAGIOCLASE

BIOTITE

QUARTZ
The nature of the zircons themselves is much more obvious in the mineral separates (Plates I, II). The zircon population is a mixture of crystals and crystal fragments. The majority of the crystals are subhedral prisms with one or two simple pyramidal terminations. All the grains have rough crystal faces. Under the light microscope it is not possible to tell whether this is due to irregular overgrowth or corrosion, but from the electron micrographs, it appears that the latter is the case (Plate II (a), (b)). The crystal faces are well developed but deeply etched. In one grain which has been broken, presumably during magma genesis, the sharp edges of the fracture clearly have been rounded (Plate II (e)).

Growth zoning is very common and cores are also, even in the finest grainsizes (Plate I (c), (f)). The cores appear to be euhedral zircon which crystallised early, not rounded detrital zircon which later has been overgrown. Inclusions are not common and, where present, are tiny transparent needles. All the zircon is cloudy and strongly magnetic (Appendix G), presumably because it has been severely damaged by radiation. This is consistent with the observed high uranium content (Table 3.1).

3.3 GEOCHRONOLOGY

From field relationships, the Maffra is known to be one of the youngest plutons in the Berridale Batholith. To within the analytical uncertainties, the ages measured by rubidium–strontium on biotite (412.8 ± 1.9 m.y.), rubidium–strontium on muscovite (413.6 ± 3.7 m.y.) and potassium–argon on biotite (412.4 ± 6.8 m.y.) are the same. The mean of these mica ages is 412.9 ± 1.6 m.y. The scarcity of accessible fresh outcrop made sampling of the Maffra for geochemistry very difficult, so only two total-rocks were analysed for rubidium–strontium. Nevertheless, sufficient dispersion in the total-rock rubidium to strontium ratio was achieved ($^{87}$Rb/$^{86}$Sr was 4.3 and 6.2 respectively) for the estimated precision of the calculated age to be relatively high. The total-rock age
Plate I  Light photomicrographs of zircons from the Maffra Adamellite.

Scale subdivisions are 10µ. The fractions illustrated are:

(a)  45 - 75µ (5)    (b)  - 45µ (3)
(c)   - 45µ (3)      (d)  + 75µ (5)
(e)   + 75µ (4)      (f)  - 45µ (3)
Plate II  Scanning electron photomicrographs of zircons from the Maffra Adamellite. Scale subdivisions are 30μ. The zircons are from fraction 45 - 75μ (5).
was 424.9 ± 9.6 (2 s.e.) m.y. This is significantly different from the mean mica age at the 5% level if it is assumed that the estimate of analytical precision takes into account all possible sources of error. Since that assumption has been proven correct in only one of the ten other plutons in the Batholith for which there are sufficient total-rock analyses for it to be tested, the difference between the total-rock and mica ages is not considered to be significant, pending further analyses.

3.3.1 Monazite

Both zircon and monazite from the Maffra were analysed. The analytical data are listed in Table 3.1 and plotted in Fig. 3.2. The monazite analysis was done in triplicate. The ages calculated from each of the three $^{207}\text{Pb}/^{235}\text{U}$, $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios are the same within error, as are the ages calculated from the weighted mean values of $^{207}\text{Pb}/^{235}\text{U}$, $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$. The three monazite analyses are indistinguishable therefore, and the mean analysis is concordant (namely, it plots within error on the concordia). These two facts are important with regard to both the accuracy of the analyses and the interpretation of the monazite data.

The three determinations were not strictly a triplicate analysis, for one sample (MM) was treated differently from the other two. MM was the first monazite to be analysed from the Berridale Batholith and was processed as a trial sample. After the sample had been dissolved, three aliquots were prepared. The first was spiked with the same mixed $^{208}\text{Pb}$-$^{235}\text{U}$ tracer as was in routine use for the zircon work, and then analysed for uranium. To the second was added a small amount of a standard solution of SRM983 (radiogenic lead) which served as a $^{206}\text{Pb}$ spike. This aliquot was analysed to determine the lead concentration. The third was used in measuring the lead composition. This procedure was adopted in order to minimise the problem of error magnification which would have arisen had a sample strongly enriched in $^{208}\text{Pb}$ (which the monazite was expected to be)
Table 3.1  Uranium-lead analytical data from the Maffra Adamellite (75-117).
The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead from potassium feldspar (Appendix E).
### Table 3.1
ANALYTICAL DATA FOR THE MAFFRA ADAMELLITE (75-117)

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight (mg)</th>
<th>Pb (ppm)</th>
<th>U (ppm)</th>
<th>(\frac{206_{\text{Pb}}}{204_{\text{Pb}}}) measured</th>
<th>Atom Percent Radiogenic Lead</th>
<th><em>Atomic Ratios</em></th>
<th>Apparent Ages (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>206\text{Pb}</td>
<td>207\text{Pb}</td>
<td>208\text{Pb}</td>
</tr>
<tr>
<td>43/ 45-75(\mu) (3)</td>
<td>2.55</td>
<td>32.23</td>
<td>780.41</td>
<td>479.7±2.2</td>
<td>81.53</td>
<td>4.68</td>
<td>13.79</td>
</tr>
<tr>
<td>45/ -45(\mu) (5)</td>
<td>4.94</td>
<td>52.99</td>
<td>840.32</td>
<td>1193±8</td>
<td>80.67</td>
<td>4.51</td>
<td>14.82</td>
</tr>
<tr>
<td>42/ +75(\mu) (5)</td>
<td>2.35</td>
<td>64.18</td>
<td>991.94</td>
<td>422.3±2.6</td>
<td>81.66</td>
<td>4.62</td>
<td>13.72</td>
</tr>
<tr>
<td>40/ 45-75(\mu) (5)</td>
<td>10.95</td>
<td>69.64</td>
<td>1129.72</td>
<td>2452±38</td>
<td>81.62</td>
<td>4.56</td>
<td>13.82</td>
</tr>
<tr>
<td>41/ +75(\mu) (4)</td>
<td>4.90</td>
<td>67.85</td>
<td>1295.40</td>
<td>598.3±2.4</td>
<td>82.38</td>
<td>4.67</td>
<td>12.95</td>
</tr>
<tr>
<td>81/ 45-75(\mu) (3)</td>
<td>1.41</td>
<td>68.72</td>
<td>1568.94</td>
<td>651.8±5.3</td>
<td>78.84</td>
<td>4.48</td>
<td>16.68</td>
</tr>
<tr>
<td>44/ -45(\mu) (3)</td>
<td>4.81</td>
<td>92.09</td>
<td>1818.86</td>
<td>356.6±1.0</td>
<td>65.67</td>
<td>3.74</td>
<td>30.59</td>
</tr>
<tr>
<td>72/ MONAZITE</td>
<td>0.72</td>
<td>1081.12</td>
<td>2489.53</td>
<td>817.1±2.6</td>
<td>13.26</td>
<td>0.73</td>
<td>86.01</td>
</tr>
<tr>
<td>67/ MONAZITE</td>
<td>0.64</td>
<td>1056.50</td>
<td>2503.31</td>
<td>824.3±4.8</td>
<td>13.74</td>
<td>0.76</td>
<td>85.50</td>
</tr>
<tr>
<td>M MONAZITE</td>
<td>13.76</td>
<td>1343.91</td>
<td>5722.77</td>
<td>1099±49</td>
<td>24.78</td>
<td>1.35</td>
<td>73.87</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.*
Fig. 3.2  A concordia plot of the uranium-lead data from the Maffra Adamellite (Table 3.1). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data. Curve B is the locus of recent episodic fractionated lead loss from a one-phase system 412 m.y. old for a $^{206}\text{Pb}$ diffusion rate a factor of 1.086 higher than the $^{207}\text{Pb}$ rate (calculated from the equations of Steiger & Wasserburg, 1966).
FIGURE 3.2

MAFFRA ADAMELLITE
(75-117)
been spiked with a lead tracer which was itself primarily $^{208}\text{Pb}$.

When the analytical data from MM had been obtained, it became clear that separate spiking of lead and uranium was not only undesirable but unnecessary. To within the analytical error, the monazite was concordant. In that case, it was imperative that the ratio of lead to uranium in the sample be determined as accurately as possible; the way to do that was to use a mixed lead-uranium tracer. As the only mixed tracer available at the time was the $^{208}\text{Pb}-^{235}\text{U}$ tracer, either a $^{206}\text{Pb}-^{235}\text{U}$ tracer had to be prepared solely for the analysis of monazite, or the problem of error magnification had to be overcome. The latter alternative was chosen.

The monazite lead was found to contain a smaller proportion of $^{208}\text{Pb}$ than had been expected (in the radiogenic component, $^{208}\text{Pb}/^{206}\text{Pb}$ was 2.98). Because $^{208}\text{Pb}/^{206}\text{Pb}$ in the mixed $^{208}\text{Pb}-^{235}\text{U}$ tracer was 458.6, the error magnification could be maintained at moderately acceptable levels (less than 1.5) if $^{208}\text{Pb}/^{206}\text{Pb}$ in the sample-tracer mixture was kept within the range 9 to 150. This was possible once the approximate concentration of lead in the sample was known.

The concentrations of lead and uranium in the other two of the replicates (67/ and 72/) were determined using the mixed $^{208}\text{Pb}-^{235}\text{U}$ tracer, and the lead to uranium isotopic ratios (corrected for blank and common lead) in the two are indistinguishable from those in MM. There is however, a significant difference between the measured lead isotopic compositions (Table 3.1). This is considered to be due to the fact that, in the preparation of 67/ and 72/ for analysis, greater precautions were taken to eliminate those monazite crystals which were corroded, clouded or impure.

The determination of the uranium to lead ratio in MM was dependent upon the correct calibration of the concentration of $^{206}\text{Pb}$ in the standard solution and of $^{235}\text{U}$ in the mixed $^{208}\text{Pb}-^{235}\text{U}$ tracer. On the other hand, the determination of the uranium to lead ratio in 67/ and 72/ depended upon the correct calibration of the ratio $^{208}\text{Pb}/^{235}\text{U}$ in the mixed
It is a verification of the accuracy of those calibrations that the measured lead to uranium ratios in the three samples are identical to within the independently assessed analytical error.

The fact that no heterogeneity in the monazite's lead to uranium ratio was detected and that the mean analysis is concordant, is important in the interpretation of the monazite data. If the monazite is considered not to be a closed uranium-lead system, the age least sensitive to recent chemical disturbance is that calculated from the mean value of $^{207}\text{Pb} / ^{206}\text{Pb}$ (418.3 ± 5.8 m.y.). This is the method of assessing monazite data which Tilton and Nicolaysen (1957), Pasteels (1970) and Doe (1970), for example, used or recommended. On the other hand, if a closed uranium-lead system is assumed, the ages calculated from $^{207}\text{Pb} / ^{235}\text{U}$, $^{206}\text{Pb} / ^{238}\text{U}$ and $^{207}\text{Pb} / ^{206}\text{Pb}$ are equally valid. In that case, the isotopic ratio least sensitive to errors in correction factors, such as blank or common lead, namely $^{206}\text{Pb} / ^{238}\text{U}$, would determine the age most precisely. This is the way monazite ages were calculated by Köppel and Grünenfelder (1971), Köppel (1974) and Hänny et al. (1975), for instance.

For the Maffra, the monazite age calculated from the mean $^{206}\text{Pb} / ^{238}\text{U}$ is 412.2 ± 2.3 m.y. This is indistinguishable from the mean mica age (412.9 ± 1.6 m.y.).

3.3.2 Zircon

In direct contrast to the analyses of monazite, the analyses of zircon from the Maffra are strongly discordant (Fig. 3.2 ). Although it would appear, on a cursory examination of the data, that the discordance is attributable to a recent net loss of lead, in detail this simple explanation is not sufficient. A linear regression analysis of the data (using the method of McIntyre et al. (1966) and correlation coefficients calculated using the approximation of Dalrymple and Lanphere (1974)) showed that other processes also had to be considered.
First, although a straight line may be fitted to within analytical error to the zircon data alone (Line A, Fig. 3.2) or to the zircon and monazite data in combination, that line cannot pass through the origin. The intercept on the $^{207}\text{Pb}/^{235}\text{U}$ axis in both cases is positive ($0.0144 \pm 0.0047$ and $0.0186 \pm 0.0037$ respectively). Secondly, if a line is forced through the zircons to intersect the concordia at the point which corresponds to the mean monazite $^{206}\text{Pb}/^{238}\text{U}$ age, a perfect fit of the data to within analytical error is not obtained. When unconstrained, the regression line for the zircon data alone intersects the concordia at $438.6 \pm 9.2$ m.y., which is significantly older than either the monazite $^{206}\text{Pb}/^{238}\text{U}$ age ($412.3 \pm 2.3$ m.y.) or $^{207}\text{Pb}/^{206}\text{Pb}$ age ($418.3 \pm 5.8$ m.y.).

There are at least three explanations for this pattern of discordance which need to be considered: the common lead in the zircons may have been more radiogenic than the lead in potassium feldspar (the composition assumed for the common lead correction (Appendix E)), the zircons may have contained a small component of older lead which was inherited, or the process by which lead loss occurred may have been mass discriminatory.

1) **Radiogenic common lead**

The magnitude of the common lead correction differs considerably from one zircon analysis to the next, but if the common lead $^{207}\text{Pb}/^{204}\text{Pb}$ is assumed to be 16.7, it is possible to obtain corrected zircon analyses which might be interpreted by net lead loss from zircon with a primary age of about 412 m.y. (Fig. 3.3). This common lead composition is most unlikely, however. First, the same common lead composition must be applied in the correction of the monazite data, and the monazite then plots discordantly above the concordia. Secondly, the data can no longer be fitted perfectly by a straight line. Independently of these arguments, the necessary $^{207}\text{Pb}/^{204}\text{Pb}$ ratio of 16.7 is so much higher than the $^{207}\text{Pb}/^{204}\text{Pb}$ ratio measured in the potassium feldspar (15.57) or in present-day ore lead (about 15.6) that a radiogenic component must be considered.
Fig 3.3  The effect on the Maffra uranium-lead data of increasing the assumed common lead $^{207}\text{Pb}/^{204}\text{Pb}$ ratio from 15.57 to 16.7. The zircon analyses no longer are fitted perfectly to a straight line and the monazite analyses are discordant.
2) **Inherited radiogenic lead**

If the discordance pattern is to be explained by lead inheritance, it is necessary that the inheritance be such that increasingly discordant fractions also contain more radiogenic lead, for a progressive increase in $^{207}\text{Pb}/^{206}\text{Pb}$ with decreasing $^{206}\text{Pb}/^{238}\text{U}$ is observed. The possible solutions are most easily visualised if the discordia is considered to be a mixing line.

Unlike the more common pattern of inheritance, in which fractions with higher $^{207}\text{Pb}/^{206}\text{Pb}$ ratios also have higher lead to uranium ratios (as documented by Köppel and Grünenfelder (1971), for instance), discordance in the Maffra cannot be modelled by the mixing of two concordant uranium-lead systems. Rather, if two-component mixing is employed, one end-member must be very strongly discordant and have a $^{206}\text{Pb}/^{238}\text{U}$ age less than or equal to the lowest $^{206}\text{Pb}/^{238}\text{U}$ age observed (215 m.y. for 44/) and a $^{207}\text{Pb}/^{206}\text{Pb}$ age greater than or equal to the highest $^{207}\text{Pb}/^{206}\text{Pb}$ age observed (505 m.y. for 43/). Such multi-component mixing with at least one strongly discordant end-member has been discussed by Steiger and Wasserburg (1966).

The alternative explanation for the correlation between discordance and inheritance is that more than two components are involved, and that the proportions in which they are mixed are in some way systematically determined. It would be possible, for example, to generate the observed pattern by the mixing of zircon 412 m.y. old, which had suffered varying degrees of recent lead loss, with zircon containing lead which has a $^{207}\text{Pb}/^{206}\text{Pb}$ ratio greater than 0.05735 (the highest observed). If this model is assumed, for any particular mixture, the relationship between the proportion of the total lead which is inherited and the composition of the inherited component may be calculated. This was done for the zircon fraction which was the most radiogenic of those measured (43/) and the results are plotted in Fig. 3.4.
Fig. 3.4 The relationship between the proportion of the total $^{207}\text{Pb}$ which is contributed by inheritance and the $^{207}\text{Pb}/^{206}\text{Pb}$ age of the inherited component for fraction 43/. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of the lead in the magmatic zircon is assumed to be 412 m.y.
LEAD INHERITANCE

in

FRACTION 43/

FIGURE 3.4
Fig. 3.4 shows the percentage of the total $^{207}\text{Pb}$ in fraction 43/ which must have been contributed by an inherited component, as a function of that component's $^{207}\text{Pb}/^{206}\text{Pb}$ age. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of the sample lead is assumed to be 412 m.y. The amounts of inheritance required are not small. If, for example, the inherited lead is assumed to have a $^{207}\text{Pb}/^{206}\text{Pb}$ ratio of 0.1114 (which is the highest $^{207}\text{Pb}/^{206}\text{Pb}$ yet measured on zircon from the local sediments and equivalent to an age of 1825 m.y.), $^{207}\text{Pb}$ contributed by the inherited component would be more than 8% of the total $^{207}\text{Pb}$. For younger inherited components, the percentage of the total $^{207}\text{Pb}$ which is inherited needs to be even larger. Although the data can be explained by multi-component mixing, with the relative proportions of the components systematically related so as to generate the observed inverse correlation between $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$, a less complex model is that of lead loss with fractionation.

3) **Lead loss with fractionation**

The Maffra discordance pattern closely resembles a pattern of discordance reported by Steiger and Wasserburg (1966). Steiger and Wasserburg found that when zircon lead-thorium-uranium data from the Sandia Granite were plotted on a concordia-type diagram, the line defined intersected the lead-uranium axis to the right of the origin. The explanation they proposed was that the sample had suffered mass discriminatory episodic or diffusive loss of lead.

Steiger and Wasserburg (1966) modified the original equations of Wetherill (1956b) to incorporate a factor which related the rates of diffusive loss of the different lead isotopes. The loci of fractionated lead loss by continuous diffusion and episodic loss differed only marginally, especially when the difference between the diffusion rates was small. The locus corresponding to recent episodic fractionated lead loss from a one-phase system 412 m.y. old is plotted in Fig. 3.2 (Curve B). The diffusion rate for $^{206}\text{Pb}$ was chosen to be a factor of 1.086 higher than that for $^{207}\text{Pb}$.
There is a close coincidence between the zircon and monazite data and the calculated curve. This does not suffice as an explanation for the cause of the fractionated lead loss but it does suggest that the zircons have a variable deficiency of lead with respect to uranium and that $^{206}$Pb is more deficient than $^{207}$Pb. This could be due to a recent episode of lead loss with fractionation, the loss of lead by continuous diffusion with fractionation or the continuous loss of one or more intermediate daughter products from the $^{238}$U decay chain.

When the phenomenon of discordant uranium-lead ages was first observed, a commonly tendered explanation was the loss of intermediate daughter products from one, or both, of the uranium decay chains (Nier, 1939; Wickman, 1942; Giletti & Kulp, 1954; Kulp et al., 1954 and Aldrich et al., 1955, for example). As more isotopic data have become available however, models involving lead loss in its many forms (for example Wetherill 1956a, b; Wetherill, 1963; Wasserburg, 1963; Ulrych, 1963; Pidgeon et al., 1966; Steiger & Wasserburg, 1966; Craig, 1968; Goldich & Mudrey, 1972 and Allègre et al., 1974) and inheritance of older lead (Grauert & Arnold, 1968; Pidgeon, 1969; Köppel & Grünénfelder, 1971; Gulson & Krogh, 1973, and numerous others) have become more popular.

Both $^{238}$U and $^{235}$U have an intermediate daughter product in their decay chains which is an isotope of the inert gas radon, $^{222}$Rn and $^{219}$Rn (sometimes called actinon) respectively. The respective half-lives are 3.8 days and 3.9 seconds, so the former has the longer mean time to escape under suitable conditions, with the consequence that the rate of production of $^{206}$Pb is slowed relative to that of $^{207}$Pb.

In 1954, workers at the Lamont Geological Observatory attempted to test the significance of radon loss in uranium-lead geochronology by a series of radon leakage experiments on uraninite, samarskite, carnotite and zircon (Giletti & Kulp, 1954). Equilibrium radon leakage from zircon at room temperature as high as 6.2% was measured and the rate doubled at 150°C. Significantly, at higher temperatures still (200°C-300°C) the rate
of leakage fell, presumably due to the initiation of recrystallisation. These experiments led Kulp et al. (1954) to conclude that radon leakage was frequently of sufficient magnitude to affect measured ages. Due to the difference in radon half-lives, they considered the $^{207}\text{Pb}/^{235}\text{U}$ age to be the most reliable.

The discovery of the systematics of discordance by Ahrens (1955a, b) raised the first doubts about radon leakage being the sole cause of discordance. Ahrens argued that a physical, rather than chemical process was necessary to generate the discordance patterns he observed, but pointed out that neither radon loss nor loss of parent or daughter nuclides in proportion to their recoil energies was a sufficient explanation. Nevertheless, Aldrich et al. (1955) were able to compensate for the discordance that they found in Precambrian zircons from South Africa and North America by a correction for radon leakage of between 8 and 16%.

Kulp & Eckelmann (1957) reassessed the importance of radon loss and concluded that because leakage under laboratory conditions would be maximal, radon leakage was not likely to be a significant factor, except possibly in the determination of the $^{207}\text{Pb}/^{206}\text{Pb}$ age of minerals younger than Cambrian. Radon leakage continued to be considered by others as one of several mechanisms which may have combined in the generation of discordance (Russell & Ahrens, 1957; Tilton et al., 1957) but it was emphasised by Aldrich and Wetherill (1958) that to affect the measurement of ages, the leakage would have to have occurred over an extended period of time; this was in contrast to the gross effects of short episodes of uranium-lead fractionation. Because experiments by Starik (1956) had shown that the rate of radon loss was not as high as was to be expected from the difference in emanation half-lives, Aldrich and Wetherill concluded that it was unlikely to have a serious effect on the ages calculated from those primary minerals used most commonly in uranium-lead geochronology.
Little serious consideration was given to radon loss in later publications. Cobb and Kulp (1961) demonstrated that continuous loss of radon from samples of organic-rich nodules ('kolm') from a Swedish shale had occurred, but the relevance of that study to zircon work was questioned by Catanzaro and Kulp (1964). Catanzaro and Kulp decided that radon and/or radium loss (both of which would affect $^{206}\text{Pb}$ production principally) were not a likely cause of the discordance they were considering (that of zircon from some North American igneous and metamorphic rocks). Their argument was based on the general agreement of the zircon $^{207}\text{Pb}/^{206}\text{Pb}$ ages with potassium-argon and rubidium-strontium ages of cogenetic minerals, the regional constancy of $^{207}\text{Pb}/^{206}\text{Pb}$ ages, despite a range in $^{207}\text{Pb}/^{235}\text{U}$ ages, and the absence of discordance patterns which defined a locus of $^{238}\text{U}$ daughter loss. In a later paper, Catanzaro (1968) dismissed intermediate daughter loss entirely as a major contributing factor in zircon discordance.

In summary, the principal argument against radon loss is that in most studies, it has been found to be an insufficient explanation of the pattern of discordance observed. Nevertheless, it has been shown to be an important factor under certain conditions (Cobb & Kulp, 1961).

Radon loss cannot be dismissed as a possible factor in the generation of the Maffra discordance pattern. The pattern can be explained by a combination of severe recent lead loss (without fractionation) and continuous preferential leakage of radon from the $^{238}\text{U}$ decay chain since the zircons formed at 412 m.y. This is preferred to an explanation involving a recent episode of lead loss with isotopic fractionation because, even though leaching experiments have shown fractionation during lead loss to be possible (Tilton, 1956), the fractionation observed was confined to $^{208}\text{Pb}$ with respect to the other lead isotopes. Loss of $^{208}\text{Pb}$ with respect to $^{207}\text{Pb}$ and $^{206}\text{Pb}$ would be sensitive to the distribution of thorium with respect to uranium in the mineral grains in a way that loss of $^{206}\text{Pb}$ relative to $^{207}\text{Pb}$ would not be.
3.4 SUMMARY

Monazite analyses from the Maffra are concordant and identical to within analytical error. The most precise estimate of the monazite age, calculated from the mean value of \( \frac{^{206}\text{Pb}}{^{238}\text{U}} \), is 412.2 ± 2.3 m.y. This age is indistinguishable from that measured by potassium-argon on biotite (412.4 ± 6.8 m.y.), rubidium-strontium on biotite (412.8 ± 1.9 m.y.) or rubidium-strontium on muscovite (413.6 ± 3.7 m.y.). As only two total-rocks have been analysed for rubidium-strontium, the difference between these ages and the total-rock age of 424.9 ± 9.6 m.y. (calculated assuming no 'geological error') is not regarded as being significant.

The zircons from the Maffra are strongly discordant and the age calculated from them depends upon the manner in which the discordance is explained. Although the pattern may be attributed to a combination of recent episodic lead loss (or uranium gain) from a uranium-lead system 438.6 ± 9.2 m.y. or 412 m.y. old, and variable inheritance of older lead (or zircon) this is thought to be unlikely.

An explanation in which the primary age of the zircons is the same as that of all the other minerals is possible and preferred. It is that by some mechanism (possibly preferential intermediate daughter loss from the \(^{238}\text{U}\) decay chain over a long period of time) some lead has been lost from the zircons and the composition of the lead remaining in the grains has been fractionated so that \(^{207}\text{Pb}\) is relatively high.

The coincidence of the ages calculated by all methods is a strong argument in favour of that age being the age of emplacement and of rapid cooling of the granitoid.
CHAPTER 4

THE DELEGATE ADAMELLITE

4.1 THE GRANITOID

The Delegate Adamellite forms an isolated pluton at the southern end of the Berridale Batholith (Fig. F.1). It intrudes sediments which presently are grouped under the general heading of 'Adaminaby Beds'. A small portion of the pluton is included in the area discussed by White et al. (1978).

The Delegate is not as felsic as the Maffra but it is similarly thought to contain only a small component of residuum from its source. The rock is even-textured and medium-grained, with well-formed biotite crystals. The potassium feldspar is pink, as it is in the Maffra, but in contrast to the latter, the plagioclase generally is not coloured. Mafic xenoliths are more abundant than they are in the Maffra and several have been collected for chemical and isotopic analysis.

4.2 THE ZIRCON

The zircon sample of the Delegate (75-120) was collected from the same locality as the geochemical sample BB46. In thin section, the zircon is seen to be concentrated strongly in the biotite, and to a lesser extent in the plagioclase, almost to the exclusion of the quartz and potassium feldspar (Fig. 4.1). The few grains that were found in potassium feldspar were larger than the average grainsize and all had rough crystal faces or embayments, presumably due to corrosion. Corrosion was also found in many of the grains included in other minerals. Less of the zircon is present in close association with opaque minerals and apatite than in the Maffra, and the majority of the grains in that association occur as inclusions in biotite. Zircon was found in plagioclase, both in the rounded, relatively altered cores, and in the fresher, outer rims of the grains.
Fig. 4.1  The distribution of zircon in the minerals of the Delegate Adamellite, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
FIGURE 4.1
Plate III  Light photomicrographs of zircons from the Delegate Adamellite.

Scale subdivisions are 10μ. The fractions illustrated are:

(a) 45 - 75μ (5)  (b) 45 - 75μ (5)
(c) 45 - 75μ (3)  (d) +100μ (5)
(e) +100μ (4)  (f) +100μ (4)
The zircon population in the Delegate resembles that in the Maffra in being a mixture of well-formed prismatic crystals and crystal fragments, many of which have one or two simple pyramidal terminations. Other features of the Delegate zircons are quite different from those in the Maffra however. A large proportion of the grains show little or no cloudiness that could be interpreted as severe radiation damage and the crystal faces are apparently free of the effects of chemical etching (Plate III (a), (b)). Neither cores nor inclusions are abundant, although a number of examples of both have been observed. In contrast to the zircons in the Maffra, many of the Delegate zircons have terminations that are not simple pyramids but show the development of many small crystal faces (Plate III (a), (d)).

4.3 GEOCHRONOLOGY

4.3.1 Rubidium-strontium and potassium-argon analyses

Total-rock rubidium-strontium analyses were the first isotopic data obtained from the Delegate. The chemical composition of the granitoid is very uniform however, and little dispersion in the rubidium to strontium ratio was found. As a consequence, the precision of the calculated total-rock age was very low. The dispersion was improved by including analyses of several xenoliths, but with the result that the calculated age became heavily dependent upon those analyses. This procedure is justified if both the granitic melt and the xenoliths originated from the same source and if the processes of magma genesis rehomogenised the source strontium isotopic composition, or at least destroyed any correlation between rubidium and radiogenic strontium which previously had evolved. Neither the granitoid and xenolith data together, nor the granitoid data alone were fitted to a straight line to within analytical error, so the calculation of a total-rock age was possible only after the cause of the excess scatter had been modelled.

In their discussion of the statistical assessment of rubidium-
strontium isochrons, McIntyre et al. (1966) proposed three different models which could be applied in the evaluation of rubidium-strontium data, if that data did not define a straight line to within independently assessed analytical error (that is, did not satisfy the criteria for their Model 1). They were:

Model 2. Geological variation of $\frac{^{87}\text{Sr}}{^{86}\text{Sr}}$ proportional to $\frac{^{87}\text{Rb}}{^{86}\text{Sr}}$ for each sample.

Model 3. Geological variation of $\frac{^{87}\text{Sr}}{^{86}\text{Sr}}$ independent of $\frac{^{87}\text{Rb}}{^{86}\text{Sr}}$ for each sample.

Model 4. A combination of Models 2 and 3.

Although Brooks et al. (1972) questioned the utility of these models, they are the best means available for the consideration of 'geological error' and have been used in this thesis for the assessment of both the rubidium-strontium total-rock data and some of the zircon uranium-lead analyses.

A test of the scatter in the combined granitoid and xenolith total-rock data from the Delegate using the method of McIntyre et al. (1966) indicated that Model 4 was the most appropriate. Using that model, the total-rock age is $452.3 \pm 14.8$ (t s.e.).

A Student's $t$ multiplier determined by the number of points in the regression has been used in the calculation of the estimate of precision. In contrast to the total-rock result, the Delegate's biotite age is $415.1 \pm 6.8$ m.y., when measured by potassium-argon, and $415.0 \pm 2.6$ m.y., when measured by rubidium-strontium. The fact that the results of these two independent techniques are in close agreement is strong evidence for this age being geologically significant. The difference between the total-rock age and the mica age can be reduced slightly by assuming a different regression model but it cannot be eliminated.

4.3.2 Uranium-lead analyses

Both zircon and monazite occur in the Delegate, so it was possible to obtain two independent measurements of the granitoid's age by
uranium-lead. The analytical data are listed in Table 4.1 and plotted on a concordia diagram in Fig. 4.2. To within the analytical errors, the monazite ages calculated from the mean ratios $^{206}\text{Pb}/^{238}\text{U}$ (426.7 ± 2.4 m.y.), $^{207}\text{Pb}/^{235}\text{U}$ (427.3 ± 3.7 m.y.) and $^{207}\text{Pb}/^{206}\text{Pb}$ (430.7 ± 19.8 m.y.) are the same. The monazite age is therefore considered to be that which is the most precisely determined, the $^{206}\text{Pb}/^{238}\text{U}$ age, 426.7 ± 2.4 m.y.

The zircon data from the Delegate are discordant but not as strongly so as those from the Maffra (compare Figs. 3.2 and 4.2). To within analytical error, they fit perfectly to a straight line which intersects the concordia in two places, at 193.3 ± 68.6 m.y. and at 495.5 ± 40.7 m.y. respectively (Fig. 4.2, line A). Neither the monazite data nor the origin can be included in the regression with the zircon data and the perfect fit be maintained. Similarly, the zircon data and the point on the concordia which corresponds to the monazite age are not perfectly fitted to a straight line.

From these relationships alone, it is not possible to determine how the zircon and monazite data should be interpreted. The zircon may have a primary age of about 495 m.y. and have suffered an episode of lead loss at about 193 m.y., which only affected the monazite partially. The zircon may have suffered recent lead loss, and the monazite also, but to a lesser extent. There may be a small component of inherited lead in the zircon and the monazite age may therefore be correct. Because of this uncertainty, an independent means of placing a limit on the pluton's emplacement age was sought.

It is possible to constrain the emplacement age of the Delegate by two methods. First, the sedimentary rocks which the pluton intrudes are fossiliferous and an estimate of their absolute age may be obtained from the stratigraphic time scale. This age is an older limit to the emplacement age of the Delegate. Secondly, the Delegate is cut by an intrusive of aplite which must be either synchronous with, or younger
Table 4.1  Uranium-lead analytical data from the Delegate Adamellite (75-120). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar from the Maffra Adamellite (Appendix E).
<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight mg</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206&lt;sup&gt;Pb&lt;/sup&gt; measured</th>
<th>ATOM PERCENT RADIOGENIC LEAD</th>
<th>*ATOMIC RATIOS</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>204&lt;sup&gt;Pb&lt;/sup&gt;</td>
<td>206&lt;sup&gt;Pb&lt;/sup&gt;</td>
<td>207&lt;sup&gt;Pb&lt;/sup&gt;</td>
<td>208&lt;sup&gt;Pb&lt;/sup&gt;</td>
</tr>
<tr>
<td>14/ 45-75μ (5)</td>
<td>20.30</td>
<td>44.37</td>
<td>678.91</td>
<td>5906±364</td>
<td>85.54</td>
<td>4.82</td>
<td>9.64</td>
</tr>
<tr>
<td>57/ 75-100μ (5)</td>
<td>5.30</td>
<td>55.66</td>
<td>820.85</td>
<td>3044±15</td>
<td>85.74</td>
<td>4.83</td>
<td>9.44</td>
</tr>
<tr>
<td>58/ +100μ (4)</td>
<td>4.44</td>
<td>68.60</td>
<td>1011.16</td>
<td>3956±41</td>
<td>84.61</td>
<td>4.74</td>
<td>10.65</td>
</tr>
<tr>
<td>15/ -45μ (3)</td>
<td>15.28</td>
<td>64.98</td>
<td>1055.79</td>
<td>3639±33</td>
<td>82.75</td>
<td>4.59</td>
<td>12.66</td>
</tr>
<tr>
<td>13/ 45-75μ (3)</td>
<td>10.82</td>
<td>82.59</td>
<td>1205.60</td>
<td>3183±393</td>
<td>79.40</td>
<td>4.51</td>
<td>16.09</td>
</tr>
<tr>
<td>84/ MONAZITE</td>
<td>0.84</td>
<td>870.37</td>
<td>1085.33</td>
<td>643.3±11.0</td>
<td>7.49</td>
<td>0.42</td>
<td>92.09</td>
</tr>
<tr>
<td>85/ MONAZITE</td>
<td>0.61</td>
<td>908.89</td>
<td>1087.22</td>
<td>493.2±8.2</td>
<td>7.16</td>
<td>0.40</td>
<td>92.44</td>
</tr>
<tr>
<td>69/ MONAZITE</td>
<td>1.57</td>
<td>877.00</td>
<td>1320.36</td>
<td>526.5±5.7</td>
<td>9.01</td>
<td>0.50</td>
<td>90.49</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.
Fig. 4.2 A concordia plot of the uranium-lead data from the Delegate Adamellite (Table 4.1). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data and intersects the concordia at 193.3 $^{+68.6}_{-91.1}$ m.y. and 495.5 $^{+40.7}_{-25.7}$ m.y. respectively. Line B is the line through the origin which best fits the zircon data. It intersects the concordia at 453.7 $^{+15.7}_{-14.5}$ m.y.
DELEGATE ADAMELLITE
(75-120)

FIGURE 4.2
than, the granitoid. The age of the aplite provides a younger limit to the granitoid's age of emplacement.

4.3.3 The Delegate Aplite

A feature of the Delegate is a substantial body of aplite, with an outcrop area of more than three square kilometres, which is present near the centre of the pluton. Whether it represents a fluid phase that was late to crystallise, or partial remelting of the granitoid after its emplacement (which is the less likely), the aplite cannot be older than the granitoid itself.

Samples of the aplite for rubidium-strontium total-rock geo-chronology were collected from two different localities (Fig. F.1). One sample was analysed routinely and the other was subdivided into two parts, using grainsize differences as a criterion. The analytical data for the Delegate Aplite are listed in Table 4.2 and plotted on an isochron diagram in Fig. 4.3. To within the analytical error, the three analyses are perfectly fitted to an isochron, from which an age of 415.2 ± 18.0 \( (t \text{ s.e.}) \) m.y. and an initial \( \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \) of 0.70729 ± 0.00587 \( (t \text{ s.e.}) \) may be calculated. The uncertainties in these results are considered to be unrealistically high however, and to reflect an incorrect choice of statistical model.

In their assessment of the use of two-error regression treatments, Brooks et al. (1972) considered the way in which the Student's \( t \) multiplier should be applied in the estimation of isochron errors. It was their conclusion that, in the case where the data which defined the isochron did so to within independently assessed experimental error, the \( t \) multiplier should be determined, not by the number of samples in the particular regression, but by the number of replicate analyses on which the estimate of analytical error was based.

The Delegate Aplite data are one such case. Because the aplite almost certainly formed from totally molten magma and is small in comparison
<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>
</tr>
</thead>
<tbody>
<tr>
<td>76-101A</td>
<td>264.9</td>
<td>25.92</td>
<td>30.02</td>
<td>0.88468</td>
</tr>
<tr>
<td>76-101B</td>
<td>293.0</td>
<td>24.26</td>
<td>35.58</td>
<td>0.91782</td>
</tr>
<tr>
<td>76-102</td>
<td>250.7</td>
<td>46.24</td>
<td>15.80</td>
<td>0.80073</td>
</tr>
</tbody>
</table>

**TABLE 4.2**

Rubidium, strontium and \(^{87}\text{Sr}/^{86}\text{Sr} \) for the Delegate Aplite
FIGURE 4.3

$R_{1} = 0.70729 \pm 0.00092$

$\delta S_{98}/S_{78}$
to the volume of most plutons, there is an independent expectation that it crystallised in a short time (geologically) and with an uniform initial strontium isotopic composition. This is borne out by the perfect fit of the data to a straight line to within analytical error; the mean square of weighted deviates (M.S.W.D.), an estimate of the residual variance of the regression based upon the scatter of the data about the line, is 0.23.

If the estimate of analytical precision is assumed to be based on a large number of replicates and the appropriate t multiplier of 2.0 is used, the uncertainties calculated previously are reduced by a factor of 6.4. The Aplite total-rock age is 415.2 ± 2.8 (2 s.e.) m.y. and the initial $^{87}\text{Sr}^{86}\text{Sr}$ is 0.70729 ± .00092 (2 s.e.). The precision estimates no longer take into account the possibility that the observed linearity of the data is a coincidence.

In view of the resistance of total-rock samples to resetting, the absence of any identified post-emplacement metamorphism and the rapidity with which the Aplite is likely to have cooled, 415.2 ± 2.8 m.y. is considered to be the Delegate Aplite's age of crystallisation.

4.3.4 The stratigraphic time scale

Because of the Delegate's isolation, limits to its age of emplacement may be determined from stratigraphic control, but not from contact relationships with other granitoids. Less than 1 km beyond the pluton's northern boundary, the rocks of the Adaminaby Beds contain poorly preserved graptolites to which Patterson (1968) has assigned an Eastonian age. None of the fossiliferous units actually is incorporated within the Delegate's narrow metamorphic aureole but Eastonian is nevertheless a maximum estimate for the pluton's age of emplacement.

500 m to the east of the Delegate, the Ordovician sequence is overlain unconformably by Silurian shallow-water sediments. The importance of establishing the presence or absence of contact metamorphic effects in those sediments has been recognised but, because of the very siliceous
lithologies which are lowermost in the Silurian section and the low grade of the metamorphism at the critical localities, the problem currently is unresolved (A.J.R. White, pers. comm.).

The use of the Delegate's sedimentary host rocks as a constraint on its age of emplacement is dependent upon the determination of the absolute time interval which corresponds to the Eastonian (or approximately, to the Caradoc). The few published studies which are relevant to the problem have been discussed by Williams et al. (1976) and further assessed, in combination with some new data, by Lanphere et al. (1977). The most important of the results are summarised below. For ease of comparison they have been recalculated, where necessary, to the decay constants which are used in the present work.

Metamorphic rocks in the aureole of the Bay of Islands ophiolite complex, Newfoundland, were dated by $^{40}\text{Ar} - ^{39}\text{Ar}$ by Dallmeyer and Williams (1975). Stratigraphic evidence cited by Stevens (1970) suggests that the complex was emplaced in the Tremadoc or Arenig (Early Ordovician). If it is assumed that the metamorphic rocks were formed during the tectonic emplacement of the ophiolite, the measured age of $469 \pm 5$ m.y. is a minimum for the Arenig.

Selective leaching experiments on drill core samples from the Great Australian (Artesian) Basin by Bofinger et al. (1968) generated rubidium-strontium isochrons which were interpreted as dating burial metamorphism. Their result becomes $473 \pm 3$ m.y. if it is adjusted by the same proportion as is necessary to equate analyses of NBS standards done in 1968 with those obtained at A.N.U. at the present time. $473 \pm 3$ m.y. was considered by Williams et al. (1976) to be a minimum estimate of the age of the base of the Eastonian.

Harris et al. (1965) reported potassium-argon biotite ages from the Bail Hill Volcanics, Scotland, which had been measured on andesite that was associated with Early Caradocian graptolitic shales. The analyses
were done at three different laboratories and the mean age was $456 \pm 11$ m.y.

The pooled potassium-argon age of six samples of sanidine and biotite from the Early Caradocian Kinnekulle Bentonite, Sweden, determined by Byström-Asklund et al. (1961), was $452 \pm 10$ m.y.

$^{40}\text{Ar}-^{39}\text{Ar}$ analyses of hornblende from 'auto-brecciated juvenile volcanic material' and 'porphyritic hornblende diorite' provided Lanphere et al. (1977) with a minimum estimate of the age of the Monograptus cyphus Zone of the Lower Silurian on Esquibel Island, Alaska. By extrapolating from their measurement of $433 \pm 3$ m.y. using sedimentation rates, Lanphere et al. (1977) calculated an age of $435-437$ m.y. for the Ordovician-Silurian boundary.

Because a sample kindly was supplied by Dr. H. Baadsgaard, it was possible to reassess the data originally published by Byström-Asklund et al. (1961) by a measurement of the rubidium-strontium mineral age of the Kinnekulle A₁ Bentonite. The analytical data for the sample are listed in Table 4.3 and plotted on an isochron diagram in Fig. 4.4. Two mineral separates were prepared, one of biotite (which was carefully purified) and the other of a concentrate of crystalline material, only some of which was sanidine, from which as much clay and biotite as possible had been removed ('Sanidine +').

The data do not define a straight line to within analytical error due to chemical disturbance (rubidium gain or loss of radiogenic strontium) of the total-rock; if the isochron is drawn through the total-rock and biotite data only, an impossibly low initial $^{87}\text{Sr}/^{86}\text{Sr}$ is indicated. It is likely that the phase which has not remained a closed rubidium-strontium system is the clay, and that the total-rock analysis is perturbed by rubidium adsorption. If that is so, the clay-contaminated 'Sanidine +' fraction must be considered to be on a mixing line for which the total-rock (or more strictly, total-rock minus biotite) is one end-member. The 'Sanidine +' has such a low rubidium to strontium ratio that
TABLE 4.3

Rubidium, strontium and $^{87}\text{Sr}/^{86}\text{Sr}$ for the Kinnekulle A$_1$ Bentonite (75-430)

<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>
</tr>
</thead>
<tbody>
<tr>
<td>Total-rock</td>
<td>164.5</td>
<td>23.80</td>
<td>20.16</td>
<td>0.81229</td>
</tr>
<tr>
<td>'Sanidine +'</td>
<td>18.79</td>
<td>323.4</td>
<td>0.1678</td>
<td>0.71338</td>
</tr>
<tr>
<td>Biotite A</td>
<td>374.2</td>
<td>20.39</td>
<td>54.81</td>
<td>1.0597</td>
</tr>
<tr>
<td>Biotite B</td>
<td>366.8</td>
<td>20.88</td>
<td>52.41</td>
<td>1.0446</td>
</tr>
</tbody>
</table>

Biotites A, B - replicate analyses.
Kinneklelle A₁ Bentonite
75-430

\[ \frac{^{87}\text{Rb}}{^{86}\text{Sr}} \]

\[ \frac{^{87}\text{Sr}}{^{86}\text{Sr}} \]

445.0 ± 3.1 m.y.

- Total Rock
- Biotite 1
- Biotite 2
- Sanidine 
  \[ R_i = 0.71232 ± 0.00010 \]

FIGURE 4.4
any correction applied to its analysis in allowance for clay contamination can be only minimal.

The biotite age of the Bentonite is best calculated by the construction of two-point isochrons linking the 'Sanidine +' fraction and each of the biotites. The ages of Biotites A and B, determined in this way, are 444.9 ± 4.4 m.y. and 445.1 ± 4.4 m.y. respectively. The mean rubidium-strontium biotite age for the Bentonite is 445.0 ± 3.1 m.y. and the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is 0.71232 ± 0.00010.

A more direct comparison with the original age determination was obtained by remeasuring the biotite's potassium-argon age. A duplicate analysis of the same biotite as had been separated for the rubidium-strontium work gave a mean age of 450.1 ± 3.7 m.y. (N.W. Tetley, pers. comm.). The difference between this result and the mean rubidium-strontium biotite age is only just significant at the 5% level. The difference may reflect the presence of analytical error which has not been accounted for, but is more likely to be due to the fact that the Bentonite biotite has not behaved as a perfect closed system. Neither the rubidium-strontium nor the potassium-argon biotite age is distinguishable from the pooled mean of the original determinations, recalculated to the same decay constants, 452 ± 10 m.y.

To summarise, the data which presently are published provide only a loose control on the absolute age of the Eastonian. The estimate of 435-437 m.y. for the Ordovician-Silurian boundary is a minimum younger limit. Ages measured on pre-Eastonian rocks, 473 ± 3 m.y. and 469 ± 5 m.y. do not establish the older limit for they also are minimum estimates, but they are compatible with the ages which have been measured on rocks which are stratigraphic equivalents of the Eastonian, 456 ± 11 m.y. and 452 ± 10 m.y., 445 ± 3 m.y. and 450 ± 4 m.y.

Because all the measured ages must be considered to be minimum estimates, their use in placing limits on the Delegate's emplacement age is subjective. The interpretation adopted here is as follows. The
Eastonian may have begun before 470 m.y. ago; it certainly had started by 450 m.y. ago. It must have ended 436 m.y. ago or earlier. This implies that the emplacement age of the Delegate is less than or equal to 460 ± 10 m.y.

4.4 INTERPRETATION OF THE GEOCHRONOLOGY

Because the different isotopic systems in the Delegate record different ages, the determination of the age of any particular event, such as emplacement, is not a simple matter. On the other hand, there is the potential to consider the magma as an evolving system, in which each isotopic technique defines a point in an extended history. If this is to be done, the correct interpretation of the zircon discordance is most important.

In very simple terms, the zircon discordance may be modelled by the mixing of two components, the ages of which are determined by the points of intersection of the discordia and concordia (Fig. 4.2 Line A), 193.3 ± 68.6 m.y. and 495.5 ± 40.7 m.y. respectively. By a model of episodic lead loss, the latter would be the primary age of the zircons and the former would be a time of an episode of lead loss or new zircon crystallisation. There are at least two arguments against this model: first, none of the other isotopic systems in the Delegate has recorded a 193.3 m.y. event and secondly, if zircon did crystallise at 495.5 m.y., it did so long before the pluton was emplaced and before any other of the isotopic systems closed.

An alternative explanation of the discordance is that the zircons are dispersed along a locus of lead loss (without fractionation). This implies that analysis 15/ (Fig. 4.2 ) is in error beyond the specified limits of uncertainty, but is preferable to a model of lead loss with fractionation (such as that applied to the Maffra data), for which even greater analytical error must be assumed. Because the data are not strongly discordant, it is impossible to make a choice between the models.
of recent episodic and continuous diffusive lead loss and the choice, in any case, has an insignificant effect on the calculated primary age of $453.7 \pm 14.5$ m.y. (the intersection with the concordia of a line fitted to the origin and the zircon data).

Although the zircon data alone are insufficient grounds for choosing either of these models in preference to the other, the model of unfractionated lead loss is favoured for two reasons: first, the primary age for the zircon which that model implies is supported by, and is in support of the rubidium-strontium total-rock age ($452.3 \pm 14.8$ (t s.e.) m.y.) and secondly, it is not necessary to postulate an otherwise unrecognised 193 m.y. event.

The question which arises from the conclusion that approximately 452 m.y. is a geologically significant age is whether that is the Delegate's emplacement age. The answer is that, even though it could be, given the stratigraphic constraints, it can be argued from the country rock metamorphism that it probably is not. The biotite, aplite and monazite ages show that raised temperatures persisted in the Delegate until as late as 415 m.y. ago or that if the pluton cooled, there was renewed activity at that time. If the emplacement age were 452 m.y. it follows that there was high heat flow in the vicinity of the Delegate, either intermittently or continuously, for a period exceeding 35 m.y. (the Maffra is an example of the rapidity with which a small pluton otherwise would cool). There is no evidence of prolonged high temperatures in the Delegate's country rock; the pluton's metamorphic aureole is less than 1 km in width and may be explained by contact effects alone. Emplacement at 452 m.y. is considered to be unlikely.

The remaining possibilities are that emplacement was recorded by the monazite, or by the biotite and aplite, or by neither of these. Because it is probable that the aplite represents the last phase of igneous activity and that the biotite ages were reset at that time, the biotite-aplite age of 415 m.y. is a minimum estimate of the age of emplacement.
On the other hand, the monazite age is likely to be a primary crystallisation age, and although in granitic melts emplacement and the beginning of crystallisation cannot be assumed to be synchronous, evidence such as that presented by Williams et al. (1976) and the Maffra data suggest that cooling after emplacement is relatively rapid. In that case, the monazite age is considered to be an older limit to the Delegate's emplacement age.

The decision that the Delegate was emplaced between the time the monazite closed (426.7 ± 2.4 m.y.) and the time the biotite and aplite total-rock closed (415.0 ± 2.6 m.y. and 415.1 ± 6.8 m.y., and 415.2 ± 2.8 m.y. respectively) requires that the granitoid's zircon uranium-lead and total-rock rubidium-strontium systems have by some means survived from an earlier phase of the granitoid magma's history. Because the pluton is felsic and mafic xenoliths are comparatively rare, this survival cannot be attributed to the large scale incorporation of unequilibrated source material. It is necessary to explain the older ages in terms of the history of the magma itself.

4.4.1 A geochronological model

Approximately 452 m.y. ago, a near-minimum melt formed from igneous source materials which previously had been plated onto the base of the crust. The melt rose into the crust, carrying with it only a small amount of solid source residuum, but soon was forced, by changing conditions of pressure, temperature and/or volatile content, to crystallise. Zircon formed and radiogenic strontium began to accumulate.

At or before 427 m.y. a further small amount of partial melting took place. Monazite which previously had crystallised was remelted and the magma became mobile, but the melting was insufficient to rehomogenise the total-rock rubidium-strontium or to affect the zircons.

At 427 m.y. monazite again crystallised and then, or a short time afterwards, but no later than 415 m.y., the magma rose into the crust without separation of the melt from the solids, and was emplaced into an Eastonian sedimentary pile.
The heat source remained active until 415 m.y., at which time a volatile-rich, felsic phase of the magma was intruded into the still warm bulk of the pluton. The additional heat was sufficient to raise the temperature above the biotite blocking temperature (about 300°C (Jäger & Zwart, 1968)). The whole intrusive then cooled rapidly and all the isotopic systems closed. Either during a recent episode or by continuous diffusion since crystallisation, the zircon suffered a small net loss of lead.

4.5 SUMMARY

The different isotopic systems in the Delegate record different ages. The rubidium-strontium total-rock age is heavily dependent upon analyses of xenoliths, for the granitoid has a comparatively uniform rubidium to strontium ratio. The total-rock data are not fitted to a straight line to within analytical error, but if Model 4 is used, an age of 452.3 ± 14.8 (t s.e.) m.y. is obtained.

The age calculated from the zircon data is dependent upon the interpretation of discordance. The preferred interpretation is that there has been a net loss of lead with respect to uranium from zircon which has a primary age of 453.7 ±15.7 ^{+}15.7 _{-14.5} m.y. It is possible, given the maximum age of the local sediments (460 ± 10 m.y.), that the pluton was emplaced at 450 m.y. but this is considered to be unlikely. The monazite age of 426.7 ± 2.4 m.y. is interpreted as an older limit to the age of emplacement.

The biotite ages measured by potassium-argon and rubidium-strontium (415.1 ± 6.8 m.y. and 415.0 ± 2.6 m.y. respectively) are indistinguishable from the rubidium-strontium total-rock age of aplite which intrudes the pluton (415.2 ± 2.8 m.y.). This age is a younger limit to the age of emplacement of the granitoid. It is possible, but not necessary, that the biotite ages were reset at the time the aplite was intruded.
CHAPTER 5

THE TARA GRANODIORITE

5.1 THE GRANITOID

The Tara Granodiorite occurs in a relatively small, elliptical pluton (23 km$^2$ in area) which intrudes, and is virtually surrounded by, the Cootralantra. A tiny stock of Tara Granodiorite (less than 1 km$^2$ in area) is present 3 km to the southwest of the main body, and also is intrusive into the Cootralantra. The petrology and geochemistry of the Tara have been discussed in detail by White et al. (1977).

The Tara is one of the most mafic I-type granitoids in the Berridale Batholith. Its mineralogy differs from that of the Maffra and the Delegate in that hornblende is present (about 2.5% of the mode). Only near the centre of the pluton, where the granitoid is slightly more felsic than elsewhere, is hornblende not found. This suggests that the pluton is weakly zoned. In hand specimen the granitoid is medium-grained and has an even texture similar to that of the Maffra and the Delegate. Plagioclase is the most abundant mineral and constitutes nearly 50% of the rock. It occurs as well-formed crystals that commonly consist of partially altered cores and multiply-zoned rims. The primary ferromagnesian mineral is biotite, but hornblende, despite its relatively low abundance, is commonly very conspicuous. It occurs as prismatic crystals that may be up to 1 cm in length. Potassium feldspar is interstitial.

The prevalence of mafic xenoliths is a feature of the Tara. All the xenoliths are strongly rounded and most are within the size range from a few to tens of centimetres in diameter. The composition of the xenoliths approximates that of a diorite or gabbroic diorite (White et al., 1977), so they contain, in general, more hornblende and biotite and less potassium feldspar and quartz than their host.

The possibility that the xenoliths represent pieces of the granitoid's country rock which were incorporated into the magma during its
ascent was considered by White et al. (1977), but subsequently rejected. First, the xenoliths comprise a coherent population with chemical characteristics that are related very closely to those of the Tara granitoid. Secondly, the composition and mineralogy of the xenoliths precludes their being modified pieces of the Cootralantra or of metasedimentary origin. Thirdly, no pieces of the Cootralantra, either unaltered or partially assimilated, have been found in the Tara.

White et al. (1977) also rejected the idea that significant amounts of the Cootralantra had been incorporated into the Tara magma and then totally assimilated. They argued that the chemical composition of the Tara is such that the total assimilation of biotite-cordierite granodiorite would give rise to a change in mineralogy from hornblende + biotite towards biotite + cordierite, in which case, biotite would occur alone. Even at the margins of the pluton, where the effects of assimilation would be greatest, no evidence of this change is observed. It was the conclusion of White et al. (1977) that the Tara has not been contaminated by country rock.

5.2 THE ZIRCON

The sample of the Tara collected for zircon (75-002) was taken from the same locality as geochemical sample BB86 (Fig. F.1). The distribution of zircon in the Tara is very different from its distribution in the Maffra or the Delegate. Rather than being concentrated preferentially in one or two mineral phases, the zircon is evenly dispersed through all the major minerals except hornblende. Independently of the host mineral, the association between zircon, and opaque minerals and/or apatite is a strong one. There is no clear correlation between the host mineral and the zircon grain size, except that the three largest grains observed were included in quartz. In the two thin sections examined, only one zircon was found included in the core of a plagioclase grain (Fig. 5.1).

The zircon grains themselves are markedly different from those in the Delegate and the Maffra. Stubby grains with length to width ratios less
Fig. 5.1  The distribution of zircon in the minerals of the Tara Granodiorite, as observed in thin section. (For further explanation, see the caption to Fig. 3.1).
FIGURE 5.1
than 2:1 are very common and grains with length to width ratios in excess of 3:1 are rare (Plate IV(a)). The majority of the zircons are perfectly formed crystals or fragments of such. Very few grains have simple pyramidal terminations. Most of the terminations are multifaceted, even those which appear, on low magnification, to be rounded. The crystal faces commonly are very well developed (Plate V(a), (c), (d)) but roughened surfaces or embayments, possibly indicative of chemical instability, also are observed (Plate V(a), (b), (e)). Inclusions are rare but many grains have obvious cores. In most cases, the cores appear to be zircon which has been rounded before being overgrown (Plate IV(b), (d), Plate V(b), (c)) but this is not without exceptions (Plate IV(c)). Several examples of parallel growth were observed (Plate IV(c)). Jocelyn and Pidgeon (1974) have suggested that parallel growth implies crystallisation in a low viscosity melt and that in magmatic granitoids, this may be induced by the presence of a high proportion of zircon xenocrysts.

5.3 GEOCHRONOLOGY

5.3.1 Rubidium-strontium and potassium-argon analyses

Because the Tara not only is of considerable interest in its own right, but also is chemically and mineralogically similar to particular plutons in others of the southeastern Australian batholiths (the New England, the Bega, the Moruya and the Kosciusko, for example), its geochemistry has been studied in detail. Each of the nine granitoid and fourteen xenolith samples collected from the pluton was analysed for total-rock rubidium-strontium.

The granitoid data alone was fitted perfectly to an isochron to within experimental error and from it an age of $422.5 \pm 19.8$ m.y. (t s.e.) was calculated. The precision of the determination was low because of the granitoid's chemical homogeneity (the samples analysed had a range in $\frac{87\text{Rb}}{86\text{Sr}}$ of only 0.73) and could not be increased by the inclusion of xenolith analyses, because to do so introduced a component of scatter which
Plate IV  Light photomicrographs of zircons from the Tara Granodiorite and Tara xenolith. Scale subdivisions are 10μ. The fractions illustrated are:

Tara Granodiorite

(a) 45 - 75μ (8)  (b) - 45 M
(c) 45 - 75μ (8)/+100μ  (d) +100μ

Tara xenolith

(e) 70 -100μ (4)  (f) +100μ (3)
Plate V  Scanning electron photomicrographs of zircons from the Tara Granodiorite. Scale subdivisions are 30\(\mu\). The zircons are from fraction 45 - 75\(\mu\) (8').
exceeded that which could be attributed to analytical error. Inclusion of
the xenolith data nevertheless did not alter the age significantly. An
analysis of the scatter suggested that Model 3 was the most appropriate and
using that model, an age of $418.9 \pm 26.1$ (t s.e.) m.y. was calculated.

Biotite from five of the granitoid samples has been analysed for
rubidium-strontium. The calculated ages are identical to within analytical
error and their mean is $412.0 \pm 1.4$ m.y. This age is not significantly
different from the mean potassium-argon biotite age determined for four of
the samples, $413.9 \pm 3.1$ m.y. The weighted mean biotite age for the pluton,
$412.3 \pm 1.3$ m.y., is not significantly different from the granitoid total-
rock age at the 5% level.

From the biotite and total-rock ages, it is possible to place an
upper limit on the time which may have elapsed between the closure of the
total-rock system and cooling to the biotite blocking temperature. Given
the precision of the age determinations, if this time exceeded 20 m.y., it
would have been resolved. By the same argument however, if the biotite and
total-rock closed at the same time and then the biotite was reset within 20
m.y., that resetting could not be detected.

In order to determine whether biotite ages in the Tara had been
reset, it was necessary to consider ages which had been measured on samples
from outside the pluton. Williams et al. (1976) argued against post-
emplacement metamorphism of the Berridale Batholith on the basis of the
unique range of biotite ages which they obtained from each granitoid. A
similar, but stronger argument can be applied to the Tara specifically.

The biotite age of the granitoid which the Tara intrudes, the
Cootralantra, has been measured by rubidium-strontium and potassium-argon,
and the results are the same to within analytical error, $420.0 \pm 2.2$ m.y.
and $416.9 \pm 3.1$ m.y. respectively. The weighted mean biotite age of the
Cootralantra, $419.0 \pm 1.8$ m.y. is significantly greater than that of
the Tara, $412.3 \pm 1.3$ m.y., at the 5% level. A single biotite sample
collected from the Cootralantra only 600 m from the Tara's western margin
(sample 3, Fig. F.2) gave ages by rubidium-strontium and potassium-argon
which were not significantly different from one another, \(411.1 \pm 2.6\) m.y. and \(413.2 \pm 6.7\) m.y. respectively, but which were significantly different from the mean biotite age of the remainder of the Cootralantra, at the 5% level. On the other hand, the age of the sample was not significantly different from the mean biotite age of the nearby Tara.

Both the biotite in the Tara and the biotite in the Cootralantra near the Tara have recorded a thermal event which did not affect biotite sampled from the Cootralantra at other localities (one of which was only 4 km from the Tara (sample 83, Fig. F.2)). There is little doubt that that event was the emplacement of the Tara. It is concluded therefore, that since local differences in biotite ages such as this are found, regional resetting has not occurred, and that the biotite age of \(412.3 \pm 1.3\) m.y. is the Tara's emplacement age. This conclusion is strongly supported by a recent determination of the potassium-argon age of hornblende from the Tara. The age, \(415.8 \pm 7.4\) m.y., is not significantly different from the mean age of the biotite (N.W. Tetley, pers. comm.).

5.3.2 Uranium-lead analyses

Uranium-lead analyses of zircon (the Tara does not contain monazite) have shown that the geochronology of the Tara is more complex than the potassium-argon and rubidium-strontium ages would indicate. The zircon data are strongly discordant (Table 5.1, Fig. 5.2), but rather than there being a single functional relationship between \(^{206}\text{Pb}/^{238}\text{U}\) and \(^{207}\text{Pb}/^{235}\text{U}\), the \(^{206}\text{Pb}/^{238}\text{U}\), \(^{207}\text{Pb}/^{235}\text{U}\) and \(^{207}\text{Pb}/^{206}\text{Pb}\) ratios which were measured for different zircon fractions show only a weak correlation with one another. In order to interpret the discordance, it was necessary to determine by what parameters the degree of discordance of particular fractions was governed.

5.3.2.1 The systematics of discordance

The first analysts to observe that a discordance pattern could be generated for a single zircon sample by the subdivision of the sample
Table 5.1 Uranium-lead analytical data from the Tara Granodiorite (75-002). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar (Appendix E).
### Table 5.1

**ANALYTICAL DATA FOR THE TARA GRANODIORITE (75-002)**

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight mg</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>$\frac{206}{204}$ Pb measured</th>
<th>$\frac{206}{235}$ Pb</th>
<th>$\frac{206}{238}$ Pb</th>
<th>$\frac{207}{235}$ Pb</th>
<th>$\frac{207}{238}$ Pb</th>
<th>$\frac{208}{238}$ Pb</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>107/ +100µ</td>
<td>0.67</td>
<td>18.60</td>
<td>266.72</td>
<td>1646±106</td>
<td>84.96</td>
<td>5.13</td>
<td>9.91</td>
<td>.067786±.000344</td>
<td>.56385±.00645</td>
<td>.060329±.000623</td>
</tr>
<tr>
<td>16/ 75-100µ</td>
<td>6.25</td>
<td>24.95</td>
<td>399.69</td>
<td>1148±22</td>
<td>84.99</td>
<td>5.07</td>
<td>9.94</td>
<td>.058911±.000398</td>
<td>.48415±.00326</td>
<td>.059605±.000270</td>
</tr>
<tr>
<td>21/ +100µ</td>
<td>3.70</td>
<td>28.61</td>
<td>400.69</td>
<td>831.3±10.5</td>
<td>84.80</td>
<td>5.23</td>
<td>9.97</td>
<td>.066360±.000332</td>
<td>.56455±.00360</td>
<td>.061701±.000246</td>
</tr>
<tr>
<td>9/ +100µ</td>
<td>4.47</td>
<td>27.75</td>
<td>402.17</td>
<td>1278±82</td>
<td>84.28</td>
<td>5.59</td>
<td>10.13</td>
<td>.064836±.000330</td>
<td>.59303±.00774</td>
<td>.066338±.00802</td>
</tr>
<tr>
<td>22/ 75-100µ</td>
<td>5.40</td>
<td>28.06</td>
<td>421.51</td>
<td>1069±23</td>
<td>85.06</td>
<td>5.05</td>
<td>9.89</td>
<td>.062739±.000315</td>
<td>.51384±.00378</td>
<td>.059400±.00321</td>
</tr>
<tr>
<td>33/ 45-75µ (9')</td>
<td>11.75</td>
<td>30.27</td>
<td>463.22</td>
<td>1033±35</td>
<td>84.43</td>
<td>4.78</td>
<td>10.79</td>
<td>.060678±.000306</td>
<td>.47347±.00498</td>
<td>.056592±.000525</td>
</tr>
<tr>
<td>7/ 45-75µ (8)</td>
<td>35.65</td>
<td>29.39</td>
<td>474.53</td>
<td>1307±66</td>
<td>84.61</td>
<td>4.71</td>
<td>10.68</td>
<td>.058165±.000298</td>
<td>.44673±.00886</td>
<td>.055704±.001071</td>
</tr>
<tr>
<td>2/ 45-75µ (7)</td>
<td>12.06</td>
<td>41.58</td>
<td>567.98</td>
<td>559.9±3.0</td>
<td>84.03</td>
<td>4.78</td>
<td>11.19</td>
<td>.064308±.000322</td>
<td>.50465±.00294</td>
<td>.056916±.000170</td>
</tr>
<tr>
<td>32/ 45-75µ (8')</td>
<td>9.35</td>
<td>34.69</td>
<td>573.81</td>
<td>498.1±3.4</td>
<td>84.26</td>
<td>4.81</td>
<td>10.93</td>
<td>.052741±.000264</td>
<td>.41485±.00266</td>
<td>.057048±.000229</td>
</tr>
<tr>
<td>31/ 45-75µ (7')</td>
<td>4.50</td>
<td>35.15</td>
<td>574.64</td>
<td>394.0±0.7</td>
<td>84.27</td>
<td>4.82</td>
<td>10.91</td>
<td>.052002±.000261</td>
<td>.41026±.00222</td>
<td>.057219±.000118</td>
</tr>
<tr>
<td>35/ -45µ M</td>
<td>3.25</td>
<td>36.97</td>
<td>617.28</td>
<td>424.9±2.4</td>
<td>83.47</td>
<td>4.69</td>
<td>11.85</td>
<td>.051172±.000256</td>
<td>.39604±.00255</td>
<td>.056132±.000227</td>
</tr>
<tr>
<td>6/ -45µ</td>
<td>10.61</td>
<td>45.55</td>
<td>692.50</td>
<td>1059±47</td>
<td>83.80</td>
<td>4.69</td>
<td>11.51</td>
<td>.060581±.000307</td>
<td>.46756±.00605</td>
<td>.055975±.000670</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.*
Fig. 5.2 A concordia plot of the uranium-lead data from the Tara Granodiorite (Table 5.1). The estimates of analytical error are two standard errors.
TARA
GRANODIORITE
(75-002)

FIGURE 5.2
into smaller fractions that were chemically and isotopically different from one another were Silver and Deutsch (1961). They found that in the zircon from a Precambrian granodiorite which they were studying, there was a systematic change in both the concentration of uranium and the concentration and composition of the associated radiogenic lead (and hence, the degree of discordance) as a function of the zircon grainsize.

In a paper written a short time later (Silver & Deutsch, 1963), additional systematic relationships were discussed. It was shown, for example, that as a function of increasing grainsize, $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ rose, while lead, thorium and uranium concentrations fell. Although this observation was not itself an explanation of the relationship, it was Silver's opinion that the parameter which principally determined the susceptibility of a zircon's uranium-lead-thorium system to disturbance was its total radioactivity (Silver, 1963a,b).

Since this work, other researchers have considered the discordance of zircon as a function of its total radioactivity or, in many cases, as a function of its uranium content. The use of uranium content as an index of total radioactivity presumes either a direct correlation between the zircon's uranium and thorium contents or such a low thorium activity that the uranium activity is dominant. The systematics which have been observed are of three types:

1) The same as that first found by Silver and Deutsch (1963), namely, higher $^{206}\text{Pb}/^{238}\text{U}$ ratios in those fractions which have lower total radioactivities. Examples and Catanzaro & Kulp (1964), Steiger & Wasserburg (1966, 1969), Pidgeon (1969), Steiger et al. (1972) and Krogh & Davis (1973).

2) The opposite relationship to the original observation. Examples are Goldich et al. (1970), Grauert & Hofmann (1973), Gulson & Krogh (1973) and Grauert et al. (1974).

3) No systematic relationship discernible. Examples are Baadsgaard (1973), Lancelot et al. (1975) and Gebauer & Grünenfelder (1976).
The zircon from the Tara is in category 1. From the data plotted in Fig. 5.3, it is evident that, analyses of fractions 2/ and 6/ aside, there is a strong inverse correlation between the $^{206}\text{Pb}/^{238}\text{U}$ age and the uranium concentration. Whether the data from 2/ and 6/ are exceptions to the pattern, or are displaced due to analytical error could only be determined by replicate analyses; one of the major advantages in using a mixed lead–uranium tracer is that the accurate determination of sample and spike weights is not critical. Therefore, although analytical procedures were adopted by which the lead–uranium ratio might be measured accurately, the estimates of the absolute concentration of lead and uranium may be in error by several percent.

Researchers also have investigated further the observation of Silver & Deutsch (1961) that successively finer zircon fractions from the one sample were progressively richer in uranium. In the great majority of studies, the original findings have been confirmed, but there are exceptions (for example Naylor et al., 1970 and Pidgeon & Aftalion, 1972). In Fig. 5.4 the grainsize range and uranium content of zircon fractions from the Tara are plotted. A very close 'normal' relationship between the two parameters is clear.

A further systematic relationship which is important in the interpretation of discordance in the Tara, but which in other studies commonly has been of secondary interest, is the direct correlation between grainsize and $^{207}\text{Pb}/^{206}\text{Pb}$ age (Fig. 5.5). It could be argued that this is a necessary consequence of the observed correlation between $^{207}\text{Pb}/^{206}\text{Pb}$ age and uranium concentration, and between uranium concentration and grainsize, but this can be shown not to be so. The data plotted in Fig. 5.6 illustrate that the relationship between $^{207}\text{Pb}/^{206}\text{Pb}$ age and uranium concentration is a much weaker one than that between $^{207}\text{Pb}/^{206}\text{Pb}$ age or uranium concentration and grainsize (Figs. 5.4, 5.5). For the first, strong relationship to be a direct consequence of the weaker latter two would require a most unlikely coincidence. It follows that, unless there is another common parameter
The relationship between the $^{206}\text{Pb}/^{238}\text{U}$ age of, and the concentration of uranium in zircon fractions from the Tara Granodiorite. Analyses of fractions 2/ and 6/ are the only exceptions to a strong inverse correlation between the two parameters. The estimates of analytical error are two standard errors.
FIGURE 5.3
Fig. 5.4 The relationship between the grainsize of, and concentration of uranium in zircon fractions from the Tara Granodiorite. There is a strong inverse correlation between the two parameters.
FIGURE 5.4

TARA GRANODIORITE
ZIRCON

Uranium (ppm)

Grainsize (µ)

200  400  600
Fig. 5.5  The relationship between the grain size and $^{207}\text{Pb}/^{206}\text{Pb}$ age of zircon fractions from the Tara Granodiorite. There is a strong correlation between the two parameters. The estimates of analytical error are two standard errors.
FIGURE 5.5

TARA GRANODIORITE ZIRCON

207\text{Pb}/206\text{Pb} age (m.y.)

Grainsize (\mu)

400

50

100

800
Fig. 5.6 The relationship between the $^{207}\text{Pb}/^{206}\text{Pb}$ age of, and concentration of uranium in zircon fractions from the Tara Granodiorite. The inverse correlation between the two parameters is a weaker one than that between $^{207}\text{Pb}/^{206}\text{Pb}$ age and grainsize (Fig. 5.5). The estimates of analytical error are two standard errors.
FIGURE 5.6
which has not been considered, $^{207}\text{Pb}/^{206}\text{Pb}$ in the zircon is a function of the physical size of the grains, rather than of their chemical composition.

In summary, the discordance pattern for the Tara fundamentally is no different from the discordance patterns commonly observed in other zircon populations. It is unusual however, in that the dependence of $^{207}\text{Pb}/^{206}\text{Pb}$ upon chemical parameters is dominated by a stronger dependence upon a physical parameter, grainsize.

If the observed dispersion in the ratio of uranium to lead is assumed to have been generated principally by the net loss of lead without isotopic fractionation (either episodically or by continuous diffusion), three important conclusions about the isotopic ages of the zircon before lead loss occurred follow (Table 5.1, Fig. 5.2):

1) Some of the zircon had a $^{206}\text{Pb}/^{238}\text{U}$ age greater than the highest presently observed ($423 \pm 4$ m.y. for $^{107}/$).

2) Some of the zircon had a $^{207}\text{Pb}/^{235}\text{U}$ age greater than the highest presently observed ($473 \pm 10$ m.y. for $^{9}/$).

3) The range in $^{207}\text{Pb}/^{206}\text{Pb}$ ages must have equalled or exceeded that presently observed ($441 \pm 80$ m.y. for $^{7}/$ to $817 \pm 50$ m.y. for $^{9}/$).

These observations suggest that a component of the Tara zircon population crystallised at least 400 m.y. before the pluton was emplaced. In view of the chemical and mineralogical constraints on the assimilation of country rock by the granitoid, this component cannot be assumed to be of accidental derivation. In an effort to isolate or to concentrate the older component, zircon from one of the Tara's xenoliths was analysed.

5.4 THE TARA XENOLITH

A typical, rounded mafic xenolith (about 30 cm in diameter) (77-038) was collected from the same locality as the Tara granitoid sample. It was a fine-grained, even-textured rock, consisting predominantly of hornblende and plagioclase. The hornblende occurred as euhedral crystals up to 2 mm in length, as aggregates of much smaller crystals and as individual
small crystals scattered through the rock. Similarly, the plagioclase occurred both as phenocrysts up to 3 mm across which commonly were multiply-zoned and as much smaller euhedral crystals in the groundmass. Quartz and potassium feldspar were interstitial and both were in places poikilitic.

5.5 ZIRCON IN THE TARA XENOLITH

In thin section, the zircon in the Tara xenolith was found, with a single exception, to occur only as inclusions in feldspar (Fig. 5.7), despite the rock's high content of ferromagnesian minerals. This is in contrast to the distribution of zircon through all the major phases (with the exception of hornblende) in the Tara granitoid. The crystal shape and grain size of the zircon in the xenolith also contrasted with that of the zircon in the granitoid. The grains were large, striated needles up to 650 μ in length (Plate IV (e), (f)). Most were broken at one or both ends. Breakage was observed both in the separated grains and in thin section but it is not possible to be sure to what extent this occurred during sample preparation. Transparent, needle-like inclusions were common but no obvious cores were found. Although some grains appeared to have corroded crystal faces, the majority did not.

5.6 URANIUM-LEAD ANALYSES FROM THE TARA XENOLITH

Rather than containing a concentration of the older zircon component, the Tara xenolith was free of it. To within analytical error, the $^{207}\text{Pb}/^{206}\text{Pb}$ age of all the zircon fractions analysed is the same (Fig. 5.8, Table 5.2) and the discordance pattern can be explained simply by the net loss of lead, during a recent episode or by continuous diffusion, from zircon with a primary age of 445.5 ± 5.8 m.y. There is no evidence of older, inherited zircon and no evidence of zircon crystallisation at the time the Tara granitoid is considered to have been emplaced (412.3 ± 1.3 m.y.). With the exception of one fraction (106/), even the measured $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ages are significantly older than 412.3 ± 1.3 m.y.

Whether the xenolith represents solid residual material from the
Fig. 5.7  The distribution of zircon in the minerals of the Tara xenolith, as observed in thin section (For further explanation, see the caption to Fig. 3.1).
HORNBLENDE
K-FELDSPAR
PLAGIOCLASE
QUARTZ

TARA
XENOLITH

100μ

FIGURE 5.7
Table 5.2 Uranium-lead analytical data from the Tara xenolith (77-038). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar from the Tara Granodiorite (Appendix E).
TABLE 5.2

ANALYTICAL DATA FOR THE TARA XENOLITH (77-038)

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight mg</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206/ 204 Pb measured</th>
<th>ATOM PERCENT RADIOGENIC LEAD</th>
<th>206/ 207 Pb</th>
<th>207/ 206 Pb</th>
<th>207/ 235 U</th>
<th>207/ 238 U</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>106/ - 45μ (2)</td>
<td>0.36</td>
<td>16.65</td>
<td>223.14</td>
<td>421.4±22.8</td>
<td>77.36</td>
<td>4.55</td>
<td>18.09</td>
<td>206/238_U</td>
<td>207/235_U</td>
<td>207/206_Pb</td>
</tr>
<tr>
<td>103/ + 100μ (3)</td>
<td>0.99</td>
<td>25.32</td>
<td>325.00</td>
<td>1608±22</td>
<td>79.44</td>
<td>4.45</td>
<td>16.11</td>
<td>0.064681±.000364</td>
<td>0.52458±.01902</td>
<td>0.058821±.002118</td>
</tr>
<tr>
<td>104/ 70 - 100μ (4)</td>
<td>2.66</td>
<td>41.37</td>
<td>570.04</td>
<td>8788±653</td>
<td>79.76</td>
<td>4.44</td>
<td>15.80</td>
<td>0.070903±.000357</td>
<td>0.54709±.00322</td>
<td>0.055962±.000177</td>
</tr>
<tr>
<td>105/ 45 - 70μ (3)</td>
<td>1.56</td>
<td>53.00</td>
<td>757.41</td>
<td>2628±20</td>
<td>83.91</td>
<td>4.69</td>
<td>11.40</td>
<td>0.067143±.000336</td>
<td>0.51520±.00286</td>
<td>0.055651±.000135</td>
</tr>
</tbody>
</table>

* Precision estimates expressed as one standard error.
Fig. 5.8 A concordia plot of the uranium-lead data from the Tara xenolith (Table 5.2). The estimates of analytical error are two standard errors.
TARA XENOLITH
(77-038)
Tara's source, or is a product of the early crystallisation of the Tara magma which subsequently has been reincorporated, the xenolith and the granitoid magma can be assumed to have shared a common history. This being so, it is very likely that at the time zircon crystallised in the xenolith, it also crystallised in the granitoid.

A re-examination of the granitoid uranium-lead data, with the zircon age of the xenolith in mind, shows that it is possible that 445.5 ± 5.8 m.y. was the last time that zircon crystallised; none of the fractions analysed has a $^{207}\text{Pb}/^{206}\text{Pb}$ age significantly less than this. Therefore, although the granitoid discordance pattern could be explained by the mixing of younger zircon, approximately 412 m.y. old and an older component, there is no evidence to suggest that the former is in fact any younger than the zircon in the xenolith (Fig. 5.9).

5.7 INTERPRETATION OF THE GEOCHRONOLOGY

The interpretation of the complex uranium-lead data from the Tara is made difficult by the pluton's otherwise apparent simplicity geochemically and geochronologically. Some aspects of the geochronology are in relatively little doubt:

a) The pluton was emplaced at 412.3 ± 1.3 m.y.

b) There is a zircon component in the granitoid, but not in the xenolith, which must have crystallised a minimum of approximately 400 m.y. before emplacement.

c) There is no evidence for zircon crystallisation later than 445.5 ± 5.8 m.y. and no evidence for a disturbance of the zircon uranium-lead system at 412 m.y., but the data do not preclude either of these interpretations.

Central to an interpretation of the geochronology is an explanation of the presence of the older zircon component. Because of the sensitivity of the pluton's chemistry to any large-scale assimilation of country rock, this component cannot be assumed to be due to 'contamination' of the magma. Similarly, the granitoid's chemistry is not compatible with derivation from a
Fig. 5.9 A concordia plot of the combined data from the Tara xenolith and the Tara granitoid. Note that all the granitoid zircon fractions have $^{207}\text{Pb}/^{206}\text{Pb}$ ages equal to or significantly greater than those of the xenolith zircon fractions.
FIGURE 5.9

TARA
XENOLITH and
GRANITOID
metasedimentary source, so the component is unlikely to be of detrital origin. The older zircon must have been present in the granitoid's igneous source. This conclusion is difficult to reconcile with the absence of an older component in the xenolith, which is thought to be an unmelted piece of that source.

The explanation of the uranium-lead data is suggested by the morphology of the zircons themselves and their distribution on the rocks. In the xenolith, the zircons are fine needles, free of cores or obvious zoning and restricted to the felsic minerals (Fig. 5.7). This implies that they are not residual from the mafic source but have crystallised in the xenolith at a later time. In the granitoid, the zircons are smaller, more complex grains with overgrown cores, growth zones and a range of crystal forms. They are distributed at random through all the major minerals except hornblende (Fig. 5.1). This is in contrast to the preferential concentration of zircon in plagioclase and/or biotite which is seen in the Maffra and the Delegate (Figs. 3.1, 4.1). This suggests that the zircon at no time has been concentrated in particular minerals or, that if it has, it subsequently has been redistributed, possibly due to a high degree of partial melting during magma genesis. In the light of these observations, the following model for the evolution of the Tara is proposed.

5.7.1 A geochronological model

820 m.y. ago or earlier, partial melting of igneous material produced magma which was emplaced into the lower crust and solidified. The resultant rock was chemically heterogeneous, either because it was formed by several different magmas which were closely related in chemical composition or because a single magma was differentiated. Zircon crystallised only in the more felsic phases.

About 445 m.y. ago, the rock was remelted. The degree of partial melting was very high, the magma which was produced was mafic and only the most mafic, refractory components in the heterogeneous source did not become
completely molten. Much of the zircon which had crystallised was dissolved, but a small amount survived and preserved, in its lead composition, a record of its primary age.

The melt separated from its source, taking with it large amounts of solid residual material, but did not rise far before it crystallised rapidly. The residual zircon grains were trapped in whatever mineral happened to grow around them. Although it had crystallised, the magma remained sufficiently hot for elements to diffuse; the total-rock rubidium-strontium system remained open and a slow recrystallisation began. Only in the solid mafic residuum were original heterogeneities in the strontium isotopic composition partially preserved. Zirconium, which had been dispersed through the melt, formed new zircon over the older nuclei and where there were no nuclei (such as the felsic minerals of the partly altered mafic residuum) new crystals grew. The only mineral in which zircon did not crystallise was hornblende.

By 445 m.y., all the free zirconium had been incorporated into zircon or other minerals and zircon crystallisation stopped. The zircon uranium-lead system closed. Strontium continued to diffuse, however, for a minimum of a further 13 m.y., and probably for longer. At about 412 m.y., there was a rise in temperature or an increase in the partial pressure of volatiles and the rock once more became partially molten. The degree of partial melting was low. The magma rose without any separation of the molten and solid phases and was emplaced into the Cootralantra Granodiorite in two separate lobes.

Both the heating and cooling associated with emplacement were rapid so the zircon, which at the time had accumulated minimal radiation damage, remained a closed system. Subsequently to emplacement, the zircon suffered a net loss of lead either by continuous diffusion or in a recent episode. Only after emplacement did the temperature finally fall below the hornblende and biotite blocking temperatures.

5.8 SUMMARY

With the exception of the zircon uranium-lead data, the
geochronology of the Tara is very simple. The rubidium-strontium analyses of granitoid total-rocks are fitted to within analytical error to an isochron which is equivalent to an age of $422.5 \pm 19.8$ (t s.e.) m.y. The ages of biotite determined by rubidium-strontium ($412.0 \pm 1.4$ m.y.) and potassium-argon ($413.9 \pm 3.1$ m.y.) are not significantly different from this, nor from the hornblende potassium-argon age ($415.8 \pm 7.4$ m.y.). The emplacement age of the Tara is considered to be $412.3 \pm 1.3$ m.y.

The discordance pattern of the zircon uranium-lead data from the Tara granitoid is complex. It can be modelled by the mixture of zircon 412 m.y. old with zircon 820 m.y. old or older but, because all the zircon found in a mafic xenolith from the pluton is considered to have a primary age of $445.5 \pm 5.8$ m.y., this age is likely also to be the age of the younger zircon component in the granitoid. Because of the tight chemical constraints on the assimilation of metasedimentary material or biotite-cordierite granodiorite by the Tara magma, the older zircon is considered to have been derived from the Tara's source.
CHAPTER 6

THE S-TYPE GRANITOIDS

As a group, the S-type granitoids pose a very different geochronological problem from those which have been discussed above. The S-type granitoids are a class of igneous rocks that have chemical and mineralogical characteristics which show that they are derived from source material which has been fractionated chemically by a sedimentary cycle. Consequently, some of the assumptions which normally are made when dating an igneous rock either no longer are valid or need to be examined closely before proceeding. It is very likely, for example, that:

1. The source was a mixture of several rock types.
2. The rocks which comprised the source were of different ages.
3. Neither the rocks which comprised the source nor the minerals which comprised them shared a common history.
4. The source at no time had been isotopically homogeneous.

These possibilities do not need to be considered in the interpretation of ages determined using isotopic systems which close at temperatures well below that of the beginning of melting, for example, potassium-argon or rubidium-strontium mineral ages. They are important factors however, in the interpretation of data from those systems which either survive magmatic temperatures (zircon, and possibly monazite, uranium-lead) or depend upon the source being isotopically homogenised, either prior to, or during magma genesis (total-rock rubidium-strontium).

Failure of the above assumptions manifests itself in different ways. In the rubidium-strontium system, total-rock and mineral ages may not agree, and the scatter of the total-rock data about a single isochron may be far in excess of experimental error (Roddick, 1974; Roddick & Compston, 1977). In the zircon uranium-lead system, strong discordance which can be modelled as the mixing of magmatic zircon and an older zircon component, as first observed in Swiss paragneisses by Grauert and Arnold (1968), may be found.
These effects are observed in the data from all the S-type granitoids. The problem in each case is to establish whether or not the age calculated from the total-rock isochron and the apparent ages interpreted from the zircon discordance pattern are geologically significant. Because there is presently no independent means of assessing this significance (Compston and Chappell (1976) did not include S-type granitoids in their study of source ages), the interpretation of the isotopic ages measured on the S-type granitoids must be based on the argument of internal consistency alone. For this reason, although the isotopic data are presented below for individual plutons, the geochronological implications are discussed for the S-type granitoids as a whole.

6.1 THE COOTRALANTRA GRANODIORITE

6.1.1 The granitoid

The Cootralantra Granodiorite is one of the most abundant rock types in the Berridale Batholith (Figs. F.1, 2.1) and has been described in detail by White et al. (1977). It crops out over an area of 300 km², and a further 100 km² is thought to underlie later cover. Detailed mapping of the margins of the intrusive and a study of the distribution of accidental country-rock xenoliths within it suggest that there may be as many as fourteen plutons of Cootralantra Granodiorite, but so far, chemical and isotopic analyses have not identified particular plutons within the composite body.

The granodiorite is a medium- to course-grained, mafic rock. Its appearance is very similar to that of the anatctic Cooma Granodiorite. The biotite is irregularly distributed and commonly aggregated as microxenoliths. Fresh cordierite is rare but pinite pseudomorphs after cordierite are ubiquitous. In the rock, there is a complete range from tiny biotite aggregates, through microxenoliths, to biotite-rich xenoliths as large as 40 cm in length which have the appearance of banded metasediment. Such xenoliths are only one of four types which White et al. (1977) have distinguished.
1) **High-aluminium cordierite-rich gneisses**  This type of xenolith is the most abundant. They are characterised chemically by being low in SiO$_2$ and high in Al$_2$O$_3$. Their mineralogy is principally biotite, cordierite and quartz, but garnet, sillimanite, anthophyllite and green spinel also may be present. They commonly are angular in cross section. They may be banded or the minerals be segregated to form a gneissic texture. The cordierite-rich gneisses are considered to be metamorphosed pieces of pelitic or psammopelitic sediment.

2) **High-calcium xenoliths**  These xenoliths contain neither cordierite nor cordierite relics but instead, abundant aggregates of actinolite, presumably after hornblende. The xenoliths are interpreted as being the remnants of calcium-rich sediment and, although they are mafic, are distinct from the mafic xenoliths which occur in the I-type granitoids.

3) **Fine-grained, even-textured xenoliths which chemically and mineralologically are similar to the host granitoid**  These xenoliths are never banded. They are considered to be the products of early crystallisation of the magma which later have been re-incorporated into the melt.

4) **Milky quartz**  Lumps of quartz, rarely larger than 10 cm in diameter are common. These are considered to be the remains of quartz veins in the source rock which were disaggregated, but not melted, during magma genesis.

6.1.2 **The zircon**

The zircon sample of the Cootralantra (75-109) was collected from the same locality as geochemical sample BB19 (Fig. F.1). In the sample, the zircon is present in the highest concentration in biotite, in lesser concentrations in quartz and plagioclase, and is virtually absent from potassium feldspar (Fig. 6.1). In plagioclase, the zircon occurs preferentially in the grains' cores. Fine-grained zircon is most common in biotite and coarse-grained zircon is most common in quartz.

Most of the zircon grains are prismatic and have a length to width
Fig. 6.1  The distribution of zircon in the minerals of the Cootralantra Granodiorite, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
FIGURE 6.1

K-FELDSPAR  BIOTITE  PINITE

PLAGIOCLASE

COOTRALANTRA  GRANODIORITE

QUARTZ  100 μ
ratio in the range 1:1 to 3:1. Many of the grains are crystal fragments (Plate VI(a)). The crystal terminations very rarely are simple pyramids; most terminations are multifaceted (Plate VI(a), (b), Plate VII(b), (c), (d)). Some grains which appear rounded under low magnification in fact have developed a large number of very small crystal faces (Plate VI(b)) but there are rare exceptions in which rounding by abrasion appears to have taken place (compare Plate VI(D) with Plates XIV, XV). The grains are corroded to different extents; the effects of corrosion may be absent (Plate VII(c)), very slight (Plate VII(b), (d)) or very strong (Plate VII (e)). Relative to zircon in the Tara or the Maffra, cores are rare. Rare examples of parallel growth were found (Plate VI(c)).

6.1.3 Geochronology

6.1.3.1 Rubidium-strontium and potassium-argon analyses

Because the field evidence for the Cootralantra being of metasedimentary derivation is strong, the intrusive's geochemistry has been studied in detail. Each of the eighteen granitoid and seventeen xenolith geochemical samples has been analysed for total-rock rubidium-strontium geochronology. The scatter of the total-rock analyses about the isochron is far in excess of the estimated experimental error, whether all the data are assessed simultaneously (M.S.W.D. = 167.6) or the data for the granitoid samples alone are considered (M.S.W.D. = 115.9). The calculated total-rock age is dependent upon which of the regression models of McIntyre et al. (1966) is used, but only strongly so when the granitoid and xenolith data are considered together in the regression analysis (Table 6.1). The analysis showed that the scatter in the granitoid data was most appropriately modelled by Model 2. Since this is the conclusion which would be predicted from the field evidence that the whole of the Cootralantra was not emplaced simultaneously, the granitoid total-rock age is considered to be $471.4 \pm 21.1$ (t s.e.) m.y.
Plate VI  Light photomicrographs of zircons from the Cootralantra Granodiorite and Cootralantra xenolith. Scale subdivisions are 10μ. The fractions illustrated are:

Cootralantra Granodiorite
(a) 45 - 75μ (3)  (b) 45 - 75μ (5)
(c) 45 - 75μ (5)  (d) 100 -150μ NM

Cootralantra xenolith
(e) Fine M           (f) Coarse NM
Plate VII  Scanning electron photomicrographs of zircons from the Cootralantra Granodiorite. Scale subdivisions are 30μ. The zircons are from fraction 45 - 75μ (5).
Table 6.1

Total-rock ages (m.y.) calculated for the Cootralantra Granodiorite

<table>
<thead>
<tr>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
<th>Model 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granitoid and xenolith data combined</td>
<td>(445.1 ± 1.7)</td>
<td>465.8 ± 16.8</td>
<td>436.6 ± 21.8</td>
</tr>
<tr>
<td>Granitoid data only</td>
<td>(473.3 ± 3.7)</td>
<td>471.4 ± 21.1</td>
<td>472.3 ± 41.8</td>
</tr>
</tbody>
</table>

Biotite ages were measured by potassium-argon and rubidium-strontium on samples of the Cootralantra collected from four widely spaced localities (Fig. F.2). For each sample, the rubidium-strontium age (Williams et al., 1976) and the potassium-argon age are the same to within analytical error. Two of the samples were collected from within the metamorphic aureoles of later plutons and in both cases the Cootralantra biotites evidently have been reset to the biotite ages of those plutons. (The sample collected within the metamorphic aureole of the Tara already has been discussed.) The other two biotite samples, which are considered to be unaffected by metamorphism, have pooled rubidium-strontium, potassium-argon ages which are the same to within analytical error, 417.4 ± 2.9 m.y. and 419.9 ± 2.2 m.y. respectively. The mean biotite age for the Cootralantra is therefore 419.0 ± 1.8 m.y. Although this is the oldest mean biotite age measured on any granitoid in the Berridale Batholith and the Cootralantra is one of the oldest intrusives geologically, it does not necessarily follow that 419.0 ± 1.8 m.y. is the Cootralantra's emplacement age.

6.1.3.2 Uranium-lead analyses

Both monazite and zircon from the Cootralantra were analysed for uranium-lead. The data are listed in Table 6.2 and plotted on a concordia diagram in Fig. 6.2. The monazite ages calculated from the mean values
Table 6.2  Uranium-lead analytical data from the Cootralantra Granodiorite (75-109). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar (Appendix E).
### TABLE 6.2
ANALYTICAL DATA FOR THE COOTRALANTRA GRANODIORITE (75-109)

| Fraction | Sample Weight mg | Pb ppm | U ppm | 206\(^{\text{Pb}}\) \* measured | 206\(^{\text{Pb}}\) \text{238}\(^{\text{U}}\) | 207\(^{\text{Pb}}\) \text{235}\(^{\text{U}}\) | 207\(^{\text{Pb}}\) \text{206}\(^{\text{Pb}}\) | 208\(^{\text{Pb}}\) \text{206}\(^{\text{Pb}}\) | 206\(^{\text{Pb}}\) \text{207}\(^{\text{Pb}}\) | 207\(^{\text{Pb}}\) \text{208}\(^{\text{Pb}}\) | APPARENT AGES (m.y.) |
|----------|------------------|--------|-------|---------------------------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|
| 78/45 - 75\(\mu\) (1) | 1.07 | 22.66 | 290.45 | 861.1+33.3 | 84.06 | 5.05 | 10.89 | 0.072842+0.000369 | 0.60342+0.00789 | 0.060081+0.00728 | 453.3 | 479.4 | 606.5 |
| 79/45 - 75\(\mu\) (5) | 1.51 | 42.74 | 494.94 | 8898+233 | 86.74 | 5.97 | 7.29 | 0.066857+0.000434 | 0.82438+0.00416 | 0.068837+0.00050 | 536.9 | 610.5 | 893.8 |
| 77/75 -100\(\mu\) | 2.86 | 45.76 | 555.12 | 2361+118 | 87.02 | 5.88 | 7.10 | 0.081583+0.000410 | 0.75968+0.00535 | 0.067535+0.000336 | 505.6 | 573.8 | 854.3 |
| 80/100-150\(\mu\) NM | 2.70 | 49.99 | 610.19 | 9048+423 | 86.96 | 5.72 | 7.32 | 0.082571+0.000413 | 0.74928+0.00390 | 0.065814+0.00095 | 511.5 | 567.8 | 800.4 |
| 91/45 - 75\(\mu\) (3) | 1.75 | 54.03 | 740.22 | 13300+3680 | 88.75 | 5.53 | 5.72 | 0.075208+0.000378 | 0.64005+0.00465 | 0.061723+0.000326 | 467.5 | 502.3 | 664.5 |
| C7. 45 - 75\(\mu\) (5) | 43.75 | 55.81 | 758.40 | 19260+2264 | 88.06 | 5.84 | 6.10 | 0.075115+0.000377 | 0.68620+0.00362 | 0.066256+0.00113 | 466.9 | 530.5 | 814.5 |
| C1. -45\(\mu\) NM | 12.45 | 58.85 | 805.30 | 9873+640 | 88.43 | 5.57 | 6.00 | 0.074741+0.000379 | 0.64934+0.00494 | 0.063010+0.000365 | 464.7 | 508.1 | 708.6 |
| 10/75 -100\(\mu\) | 9.16 | 60.55 | 826.66 | 12200+7050 | 88.64 | 5.80 | 5.56 | 0.075198+0.000381 | 0.67846+0.00840 | 0.065430+0.000743 | 467.4 | 525.8 | 788.4 |
| C2. -45\(\mu\) M | 1.34 | 105.80 | 992.39 | 1650+59 | 54.13 | 3.16 | 42.71 | 0.065570+0.000333 | 0.52688+0.00439 | 0.058279+0.000393 | 409.4 | 429.7 | 540.3 |
| 5/45 - 75\(\mu\) (3) | 15.44 | 66.58 | 1036.51 | 6950+273 | 89.48 | 5.44 | 5.08 | 0.066275+0.000331 | 0.55542+0.00291 | 0.060781+0.00093 | 413.7 | 448.5 | 631.5 |
| B3/ MONAZITE | 0.44 | 1021.14 | 1053.95 | 131.7+0.7 | 6.36 | 0.35 | 93.29 | 0.068398+0.000347 | 0.51271+0.00763 | 0.054366+0.000764 | 426.5 | 420.3 | 386.2 |
| B2/ MONAZITE | 0.20 | 888.63 | 1067.91 | 154.7+0.7 | 7.42 | 0.41 | 92.17 | 0.068880+0.000351 | 0.52541+0.00593 | 0.055323+0.000562 | 429.4 | 428.8 | 425.3 |

*Precision estimates expressed as one standard error.
Fig. 6.2  A concordia plot of the uranium-lead data from the Cootralantra Granodiorite (Table 6.2). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data and intersects the concordia at 359.4 $^{+31.9}_{-60.4}$ m.y. and 1343 $^{+213}_{-228}$ m.y. respectively.
FIGURE 6.2

COOTRALANTRA GRANODIORITE (75-109)

ZIRCON

MONAZITE

1500
1000
500

B2M
83M
C2
10, C7

+ ZIRCON

X MONAZITE

207Pb/235U

207Pb/235U

0.5 0.6 0.7 0.8 0.9

0.1 0.2 0.3

1.0 2.0 3.0 4.0 5.0
of $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ are the same to within analytical error (427.9 ± 3.0 m.y., 425.6 ± 6.2 m.y. and 411.7 ± 36.4 m.y. respectively) so the best estimate of the Cootalantra's monazite age is considered to be the $^{206}\text{Pb}/^{238}\text{U}$ age, 427.9 ± 3.0 m.y. The difference between this age and the mean biotite age of the Cootalantra is significant at the 1% level.

The zircon data from the Cootalantra are strongly discordant. All the analysed fractions have $^{207}\text{Pb}/^{206}\text{Pb}$ ages significantly older than either the biotite or the monazite age. The range of $^{207}\text{Pb}/^{206}\text{Pb}$ ages is very large, 540.3 ± 29.5 m.y. to 893.3 ± 3.0 m.y. As even the youngest $^{207}\text{Pb}/^{206}\text{Pb}$ age far exceeds the estimated age of the Cootalantra's country rock (460 ± 10 m.y., see section 4.3.4), the discordance pattern is attributed to the effects of inheritance.

The simplest interpretation of discordance due to inheritance is that the discordia is a mixing line. The end-members are zircon which crystallised from the magma, and zircon from one or more older sources. The regression analysis of McIntyre et al. (1966) can be used to test whether or not the zircon data are fitted to within analytical error to a straight line. At the same time, the nature of the excess scatter, if any is detected, can be assessed. This analysis was done for the zircon data from the Cootalantra and from the other S-type granitoids.

Because the version of the McIntyre et al. (1966) program which was available was not designed for the assessment of lead data, it had to be used in a non-standard way. The $^{206}\text{Pb}/^{238}\text{U}$ axis was taken to be the x-axis so that a constant coefficient of variation in $^{206}\text{Pb}/^{238}\text{U}$ could easily be nominated. The error in $^{207}\text{Pb}/^{235}\text{U}$ was stipulated for each point. Correlation coefficients were calculated for each point using the same formula as used by Dalrymple and Lanphere (1974). So that the scatter could be analysed, the regression was done once, and then the $^{207}\text{Pb}/^{235}\text{U}$ axis was transferred to the $^{206}\text{Pb}/^{238}\text{U}$ value of the point of intersection of the concordia and the calculated discordia. It was possible then to repeat the regression and to test whether or not the scatter increased as a function of the points'
increasing distance from the discordia's lower intercept.

The regression analysis of the zircon data from the Cootralantra Granodiorite showed that the points could not be fitted to a straight line to within analytical error (M.S.W.D. = 17.4) and that the excess scatter is most appropriately described by Model 3. Namely, there was no evidence that the observed discordance is due to the mixing of a single, younger component and several older ones. From the intercepts of the calculated discordia (using Model 3) and the concordia, apparent ages of 359.4 ±31.9
-60.4 m.y. and 1343 ±213 m.y. (Fig. 6.2, Line A) were calculated. The difference between the former age and both the monazite age and the biotite age is significant at the 1% level. This suggests that 359.4 ±31.9 -60.4 m.y. is unlikely to have geological significance and it follows that the discordance pattern must be governed by more than simple mixing of magmatic zircon and an inherited component.

6.1.4 The Cootralantra xenolith

An attempt was made to concentrate or to isolate the zircon component which the Cootralantra granitoid had inherited from its source material by the separation of zircon from a cognate metasedimentary xenolith. The xenolith selected (77-039) was an angular piece of banded cordierite gneiss about 15 cm in diameter, and 5 cm thick. It was collected from the same locality as geochemical sample BB83 (Fig. F.1). The xenolith was fine-grained and the minerals were segregated into gneissic layers containing either principally biotite + altered cordierite or quartz + potassium feldspar + plagioclase.

6.1.5 Zircon in the Cootralantra xenolith

In thin section, the zircon was found to be located principally in plagioclase (Fig. 6.3). Although a large number of zircon grains also were present in pinitised cordierite, they were so fine (grainsize less than 10µ) that they represented only a very small fraction of the total zircon population. The average grainsize of the zircon in the xenolith was much
Fig. 6.3 The distribution of zircon in the minerals of the Cootralantra xenolith, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
FIGURE 6.3
less than that of the zircon in the Cootralantra granitoid. In contrast to the granitoid, there was no preferential concentration of zircon in biotite and no zircon was found in quartz.

The zircons themselves were very similar morphologically to those in the granitoid (Plate VI(e), (f)). The grains were in general, euhedral prisms and commonly had multifaceted terminations. Most had a length to width ratio in the range 2:1 to 4:1. Despite the small grainsize, several grains were found which appeared to contain a zircon core. Although some corrosion of crystal faces was observed, no examples of rounding due to abrasion were found.

6.1.6 Uranium-lead analyses from the Cootralantra xenolith

Uranium-lead analyses of zircon from the Cootralantra xenolith are listed in Table 6.3 and plotted in Fig. 6.4. The fractions analysed, in general, are slightly more enriched in radiogenic lead than those from the granitoid (Fig. 6.5), but this may well be only a consequence of a marginally greater success in concentrating the older zircon components in those fractions when the sample was subdivided in preparation for analysis. The range in the measured $^{207}\text{Pb}/^{206}\text{Pb}$ ages is $826.9^{+44.0}_{-45.2}$ m.y. to $1163.1 \pm 4.5$ m.y. These results show that the xenoliths is not a repository for a single, older end-member of the granitoid zircon population but that instead, it contains a similar mixture of older and younger zircon to that in the granitoid. This is in contrast to the Tara xenolith in which the older component was found to be absent (section 5.6).

As with the granitoid zircon data, the xenolith data are not fitted to a straight line to within analytical error (M.S.W.D. = 18.4) so the excess scatter must be modelled before a line of best fit is calculated. A test of the scatter revealed that it increases with increasing $^{206}\text{Pb}/^{238}\text{U}$ so that Model 2 is the most appropriate. The discordia calculated using this model intersects the concordia at $376.6^{+32.1}_{-75.6}$ m.y. and $1792^{+336}_{-407}$ m.y. (Fig. 6.4, Line A). Because of the enormous uncertainties, neither of these ages
Table 6.3  Uranium-lead analytical data from the Cootralantra xenolith (77-039). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar from the Cootralantra Granodiorite (Appendix E).
## TABLE 6.3
ANALYTICAL DATA FOR THE COOTRALANTRA XENOLITH (77-039)

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206(^{\text{Pb}}) (^{204}_{	ext{Pb}}) measured</th>
<th>(^{206}<em>{\text{Pb}}) (^{207}</em>{\text{Pb}}) (^{208}_{\text{Pb}})</th>
<th>(^{206}<em>{\text{Pb}}) (^{235}</em>{\text{U}})</th>
<th>(^{207}<em>{\text{Pb}}) (^{235}</em>{\text{U}})</th>
<th>(^{207}<em>{\text{Pb}}) (^{206}</em>{\text{Pb}})</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>96/ Fine NM</td>
<td>1.31</td>
<td>38.04</td>
<td>452.81</td>
<td>3752+131</td>
<td>86.34  6.07  7.59</td>
<td>.083827+.000421</td>
<td>.81225+.00593</td>
<td>.070276+.000375</td>
<td>518.9 603.7 936.4</td>
</tr>
<tr>
<td>95/ Coarse M</td>
<td>0.58</td>
<td>39.34</td>
<td>523.49</td>
<td>2608+307</td>
<td>86.58  5.77  7.65</td>
<td>.074305+.000377</td>
<td>.68287+.00804</td>
<td>.066653+.000713</td>
<td>462.0 528.5 826.9</td>
</tr>
<tr>
<td>108/ Coarse NM</td>
<td>0.80</td>
<td>42.93</td>
<td>528.22</td>
<td>3061+156</td>
<td>86.18  6.15  7.67</td>
<td>.080847+.000407</td>
<td>.79529+.00496</td>
<td>.071344+.000269</td>
<td>501.2 594.2 967.3</td>
</tr>
<tr>
<td>93/ Fine M</td>
<td>0.72</td>
<td>39.02</td>
<td>537.65</td>
<td>2424+79</td>
<td>86.98  5.80  7.22</td>
<td>.073117+.000366</td>
<td>.67204+.00413</td>
<td>.066661+.000239</td>
<td>454.9 522.0 827.2</td>
</tr>
<tr>
<td>94/ Coarse NM</td>
<td>1.39</td>
<td>48.13</td>
<td>540.40</td>
<td>3956+47</td>
<td>85.80  6.75  7.45</td>
<td>.088276+.000441</td>
<td>.95717+.00491</td>
<td>.078640+.00091</td>
<td>545.3 681.8 1163.1</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.*
Fig. 6.4  A concordia plot of the uranium-lead data from the Cootralantra xenolith (Table 6.3). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data and intersects the concordia at 376.6 \( +32.1 \) m.y. and 1792 \( +336 -407 \) m.y. respectively.
FIGURE 6.4

COOTRALANTRA
XENOLITH
(77-039)
Fig. 6.5  A concordia plot of the combined data from the Cootralantra xenolith and the Cootralantra Granodiorite. Note the generally higher $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the xenolith zircon fractions and the close similarity of the discordance patterns for the two samples.
COOTRALANTRA
XENOLITH and
GRANITOID

FIGURE 6.5
is significantly different from the corresponding age calculated from zircon data from the Cootralantra granitoid.

6.2 THE BULLENBALONG GRANODIORITE

6.2.1 The granitoid

The Bullenbalong Granodiorite, which has been described in detail by White et al. (1977), is similar to the Cootralantra Granodiorite chemically and mineralogically and crops out less than 10 km from the latter, but it is part of the Kosciusko, not the Berridale Batholith (Fig. F.1). As with the Cootralantra, the Bullenbalong granitoid has a very irregular texture and contains an abundance of metasedimentary xenoliths in a wide range of sizes. Its texture differs from that of the Cootralantra in that the minerals are much more deformed. In hand specimen, the quartz has a distinctive blue tint (Dietrich (1971) has interpreted the blue coloration of quartz as being due to stress-induced microfracturing) and the larger biotite grains commonly are kinked.

In thin section, the deformation textures are more pronounced. The quartz has undulose extinction and is recrystallised along the grain boundaries. The biotite has well-developed kink bands and fine biotite, which also is probably the product of recrystallisation, occurs at the margins of some of the larger grains.

6.2.2 The zircon

The zircon sample from the Bullenbalong (77-264) was collected from the same locality as geochemical sample KB12 (Fig. F.1). The similarity of the Bullenbalong to the Cootralantra extends also to the distribution of zircon (Fig. 6.6). The zircon occurs mostly as inclusions in biotite, with lesser concentrations of zircon in quartz and plagioclase. No zircon was found in potassium feldspar. The tendency of the largest zircon grains to be included in quartz is not as obvious as it is in the Cootralantra and no preferential concentration of zircon in the cores of plagioclase grains was
Fig. 6.6 The distribution of zircon in the minerals of the Bullenbalong Granodiorite, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
K-FELDSPAR

BIOTITE

PLAGIOCLASE

BULLENBALONG GRANODIORITE

QUARTZ

FIGURE 6.6
The zircons themselves are stubby prismatic crystals, mostly with length to width ratios in the range 1:1 to 4:1. Even on the finest grains the terminations are multifaceted (Plate VIII(a), (b)). Most of the grains are euhedral crystals or fragments of the same. Some rounding was observed but it appeared to be a chemical effect rather than due to abrasion (for example, Plate VIII(e)). In most cases, grains which appear rounded in fact are terminated by a large number of small crystal faces (Plate VIII (d)). Neither inclusions nor cores are common, but examples of both were found (Plate VIII(f)). Rarely, grains which were not simple, single crystals were observed (Plate VIII(e)).

6.2.3 Geochronology

6.2.3.1 Rubidium-strontium analyses

Because the Bullenbalong was not included in Compston et al.'s (1977a) study of the total-rock rubidium-strontium geochronology of the Berridale Batholith, there are insufficient total-rock rubidium-strontium analyses available from the pluton at present to calculate a total-rock age of sufficient precision to have a bearing on the current discussion. The pluton's biotite rubidium-strontium age, based on a single analysis, is $411.9 \pm 3.7$ m.y. This age is not detectably different from the biotite rubidium-strontium ages measured on other Kosciusko Batholith granitoids in the vicinity, but its relationship to the age of emplacement of the Bullenbalong is uncertain. The foliation of the granitoids clearly has involved recrystallisation of the micas and all the biotite ages may have been disturbed, if not totally reset, at the time the foliation was imposed. Although emplacement of the Bullenbalong and the imposition of the foliation may have been synchronous events, more work is required before this can be assumed.
Plate VIII  Light photomicrographs of zircons from the Bullenbalong Granodiorite. Scale subdivisions are 10µ. The fractions illustrated are:

(a)  - 30µ M       (b)  - 30µ NM
(c)  45 - 85µ (4)    (d)  45 - 85µ (1)
(e)  45 - 85µ (1)    (f)  45 - 85µ (1)
6.2.3.2 Uranium-lead analyses

One of the principal reasons that the Bullenbalong was selected for uranium-lead analysis was that it was the most zircon-rich granitoid in the region, and therefore was particularly suitable for use as an experimental sample in the development of mineral separation and chemical techniques. Because it was an experimental sample, the zircon uranium-lead data which were obtained are relatively imprecise.

Both monazite and zircon from the Bullenbalong were analysed. The data are listed in Table 6.4 and plotted on a concordia diagram in Fig. 6.7. The apparent ages of the monazite calculated from the mean values of $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{238}\text{Pb}$ (431.4 ± 3.1 m.y., 434.4 ± 8.1 m.y. and 449.1 ± 6.7 m.y. respectively) are the same to within analytical error so the monazite age is considered to be 431.4 ± 3.1 m.y. This is significantly different from the biotite rubidium-strontium age at the 1% level.

As for the Cootralantra, the zircon discordance pattern for the Bullenbalong is dominated by the effects of inheritance (the $^{207}\text{Pb}/^{206}\text{Pb}$ age measured on different fraction ranges from 593.4 ± 6.2 m.y. to 997.0 ± 25.3 m.y.). The fact that a regression analysis of the points shows that they are fitted to a straight line to within analytical error is attributable to the large uncertainties in the data and is not considered necessarily to be significant in the interpretation of the discordance. The intercepts of the calculated discordia and the concordia are equivalent to ages of 412.5 ± 11.1 m.y. and 1904 ± 132 m.y. respectively (Fig. 6.7, Line A). The former age is significantly different from the monazite age, but is not significantly different from the biotite age at the 5% level.

6.3 THE NUMBLA VALE ADAMELLITE

6.3.1 The granitoid

The most felsic S-type granitoid in the Berridale Batholith is the Numbla Vale Adamellite. It has been described in detail by White et al.,
Table 6.4  Uranium-lead analytical data from the Bullenbalong Granodiorite (77-264). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar from the Cootralantra Granodiorite (Appendix E).
TABLE 6.4
ANALYTICAL DATA FOR THE BULLENBALONG GRANODIORITE (77-264)

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight mg</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206_Pb* 204_Pb* measured</th>
<th>ATOM PERCENT RADIOGENIC LEAD</th>
<th>206_Pb 238_U 207_Pb 235_U 207_Pb 206_Pb</th>
<th>206_Pb 238_U 207_Pb 235_U 207_Pb 206_Pb</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B2. 45 - 85μ (1)</td>
<td>31.7</td>
<td>54.25</td>
<td>638.77</td>
<td>3168±258</td>
<td>86.19 6.24 7.57</td>
<td>.083898+.000450 .83742+.00652 .072392+.000454</td>
<td>519.3 617.7 997.0</td>
<td></td>
</tr>
<tr>
<td>B4. -30μ NM</td>
<td>30.4</td>
<td>54.04</td>
<td>649.77</td>
<td>3131±98</td>
<td>86.57 6.04 7.39</td>
<td>.082540+.000414 .79382+.00437 .069753+.000164</td>
<td>511.3 593.3 921.1</td>
<td></td>
</tr>
<tr>
<td>B3. -30μ M</td>
<td>30.2</td>
<td>92.21</td>
<td>1029.57</td>
<td>804.1±22.8</td>
<td>70.95 4.24 24.81</td>
<td>.069495+.000659 .57223+.00777 .059719+.000893</td>
<td>433.1 459.5 593.4</td>
<td></td>
</tr>
<tr>
<td>B5. 45 - 85μ (4)</td>
<td>27.6</td>
<td>79.09</td>
<td>1095.18</td>
<td>1824±59</td>
<td>88.47 5.42 6.11</td>
<td>.072087+.000392 .60880+.00434 .061252+.000339</td>
<td>448.7 482.8 648.1</td>
<td></td>
</tr>
<tr>
<td>110/ MONAZITE</td>
<td>0.42</td>
<td>1107.83</td>
<td>1055.86</td>
<td>238.9±3.4</td>
<td>5.81 0.33 93.86</td>
<td>.069128+.000355 .53365+.01001 .055990+.001015</td>
<td>430.9 434.2 452.0</td>
<td></td>
</tr>
<tr>
<td>109/ MONAZITE</td>
<td>0.78</td>
<td>1071.76</td>
<td>1084.22</td>
<td>279.4±3.4</td>
<td>6.18 0.35 93.47</td>
<td>.069306+.000352 .53395+.00772 .055876+.000759</td>
<td>432.0 434.4 447.4</td>
<td></td>
</tr>
</tbody>
</table>

* Precision estimates expressed as one standard error.
Fig. 6.7 A concordia plot of the uranium-lead data from the Bullenbalong Granodiorite (Table 6.4). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data and intersects the concordia at $412.5^{+11.1}_{-15.1}$ m.y. and $1904^{+132}_{-143}$ m.y. respectively.
BULLENBALONG
GRANODIORITE
(77-264)
(1978). The pluton is deeply weathered and very poorly exposed, but is thought to underlie an area of about 63 km² (Fig. F.1). The exposure is so poor that in some places the pluton was mapped only by the lack of outcrop of its country rock.

The Numbla Vale is demonstrably younger than the adjacent Finister Granodiorite. The boundary between the two plutons is convex towards the Finister, and a number of accidental xenoliths of the Finister granitoid occur in the Numbla Vale. These xenoliths have been cited by White et al. (1978) as evidence that the Numbla Vale may itself be two adjacent plutons. The position of the proposed boundary between the two is marked in Fig. F.1.

In hand specimen, the Numbla Vale appears to be moderately coarse-grained, but the grain size in fact is only 3-4 mm. The 'coarse-grained' texture is due to the strong tendency of some of the minerals, for example biotite and quartz, to occur in aggregates. The granitoid is very close to minimum melt composition (Chappell & White, 1976) and is virtually free of mafic xenoliths, but it contains abundant euhedral cordierite and cordierite pseudomorphs. These may be remnant crystals from the source material (as suggested for New England Batholith granitoids by Flood and Shaw (1975)) but it is considered more likely that they have crystallised from the magma (as suggested for the Tingaringy Granodiorite (Fig. F.1) by White et al. (1978)).

These observations imply that the Numbla Vale magma probably contained very little solid residual source material.

Although the Numbla Vale is very felsic, many of the plagioclase grains have a calcic core, as in the plagioclase in the other S-type granitoids. In other respects the texture of the rock in thin section is atypical. The potassium feldspar occurs as large, irregular crystals and myrmekite outgrowths from plagioclase into potassium feldspar are common. The aggregates of quartz are either large, partly polygonised grains with undulose extinction and deformation lamellae or a mass of small crystals, commonly with triple junctions.
6.3.2 The zircon

The zircon sample from the Numbla Vale (75-107) was collected from the same locality as geochemical sample BB2 (Fig. F.1). The distribution of zircon in the Numbla Vale differs slightly from that in the other S-type granitoids discussed above (Fig. 6.8). There is only a weak preferential concentration of zircon in biotite and plagioclase and some zircon was found in potassium feldspar. Zircon of different grainsizes is distributed relatively evenly between the principal minerals.

In thin section, several examples of zircon grains grown together were observed (Fig. 6.8) but in the grain amounts such examples were found to be rare, presumably due to disaggregation of the grains during the sample preparation. The majority of the zircons are euhedral, prismatic crystals with length to width ratios in the range 1:1 to 4:1. Although multifaceted terminations are common, simple pyramidal terminations also are found (Plate IX(a), (b), (c)). Inclusions in the grains are rare, and several examples were found in which the inclusions appeared to be non-crystalline (Plate IX(e)). Relatively few examples of zircon cores were observed (Plate IX(f)). Most of the grains show no evidence of chemical corrosion.

6.3.3 Geochronology

6.3.3.1 Rubidium-strontium and potassium-argon analyses

Because of the difficulty in obtaining fresh samples from the Numbla Vale, only three total-rocks have been analysed for rubidium-strontium. The data are perfectly fitted to an isochron to within analytical error. With so few samples, the probability that this goodness of fit is coincidental is high, but it is nevertheless thought to be significant for two reasons. First the $^{87}\text{Rb}/^{86}\text{Sr}$ ratios of the three samples are evenly dispersed over the range 3.6 to 7.1. Secondly, because the chemical composition of the granitoid is close to that of a minimum melt, it is very likely that the strontium isotopes in the magma were at some time in equilibrium. It is considered that the most appropriate t multiplier for the precision estimates is 2.0,
Fig. 6.8 The distribution of zircon in the minerals of the Numbla Vale Adamellite, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
Plate IX  Light photomicrographs of zircons from the Numbla Vale Adamellite.

Scale subdivisions are 10μ. The fractions illustrate are:

(a)  - 45μ (3)       (b)  45 – 75μ (5)
(c)  45 – 75μ (5)     (d)  45 – 75μ (3)
(e)  +100μ (4)        (f)  75 –100μ (5)
in which case the total-rock age of the Numbla Vale is $414.7 \pm 5.4$ (2 s.e.) m.y.

Biotite ages were measured for two of the samples by rubidium-strontium and potassium-argon. The difference between the ages obtained by potassium-argon ($422.2 \pm 4.9$ m.y. and $414.9 \pm 4.8$ m.y.) is significant at the 5% level, but the difference between the ages obtained by rubidium-strontium from the same samples ($417.5 \pm 2.2$ m.y. and $418.5 \pm 3.7$ m.y respectively) is not. Because the potassium-argon age and the rubidium-strontium age for each sample is the same to within analytical error, the difference between the two potassium-argon ages is considered to be due to random error. The mean biotite age for the Numbla Vale (all the data included) is $417.9 \pm 1.7$ m.y. This is not detectably different from the total-rock age.

6.3.3.2 Uranium-lead analyses

Both zircon and monazite from the Numbla Vale were analysed for uranium-lead. The data are listed in Table 6.5 and plotted in Fig. 6.9. The monazite was analysed in duplicate and the two estimates of $^{206}\text{Pb}/^{238}\text{U}$ obtained are the same to within analytical error. However, the two estimates of $^{207}\text{Pb}/^{235}\text{U}$ are significantly different and the two estimates of $^{207}\text{Pb}/^{206}\text{Pb}$ are significantly different at the 5% level. In view of the duplication of the value of $^{206}\text{Pb}/^{238}\text{U}$, these differences are considered to be due to an underestimation of the analytical error in the determination of $^{207}\text{Pb}$. This conclusion is supported by the fact that the ages calculated from the mean values of $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ are the same to within analytical error ($439.8 \pm 3.0$ m.y., $438.7 \pm 5.7$ m.y. and $436.3 \pm 21.3$ m.y. respectively). The monazite age of the Numbla Vale is considered to be $439.8 \pm 3.0$ m.y.

The zircon data from the Numbla Vale show a similar pattern of discordance to the zircon data from the other S-type granitoids, but the discordance is not as strong. The range in measured $^{207}\text{Pb}/^{206}\text{Pb}$ ages is
Table 6.5  Uranium-lead analytical data from the Numbla Vale Adamellite (75-107). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar (Appendix E).
**TABLE 6.5**

**ANALYTICAL DATA FOR THE NUMBLA VALE ADAMELITE (75-107)**

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight mg</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206 Pb *&lt;sub&gt;204&lt;/sub&gt; measured</th>
<th>206 Pb</th>
<th>207 Pb</th>
<th>208 Pb</th>
<th>206 Pb</th>
<th>207 Pb</th>
<th>208 Pb</th>
<th>238 U</th>
<th>235 U</th>
<th>206 Pb</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>27/ +100μ (4)</td>
<td>7.50</td>
<td>61.17</td>
<td>816.24</td>
<td>9255+92</td>
<td>87.58</td>
<td>5.71</td>
<td>6.71</td>
<td>0.075990±0.000380</td>
<td>0.68335±0.00342</td>
<td>0.065221±0.000078</td>
<td>472.1</td>
<td>528.8</td>
<td>781.5</td>
<td></td>
</tr>
<tr>
<td>17/ 75-100μ (5)</td>
<td>10.75</td>
<td>65.52</td>
<td>859.37</td>
<td>6977+26</td>
<td>87.59</td>
<td>5.68</td>
<td>6.73</td>
<td>0.077072±0.000385</td>
<td>0.68856±0.00345</td>
<td>0.064796±0.00019</td>
<td>478.6</td>
<td>531.9</td>
<td>767.7</td>
<td></td>
</tr>
<tr>
<td>19/ 45-75μ (5)</td>
<td>7.20</td>
<td>69.58</td>
<td>939.38</td>
<td>9234+430</td>
<td>87.51</td>
<td>5.62</td>
<td>6.87</td>
<td>0.075007±0.000375</td>
<td>0.66440±0.00346</td>
<td>0.064243±0.00096</td>
<td>466.3</td>
<td>517.3</td>
<td>749.6</td>
<td></td>
</tr>
<tr>
<td>20/ -45μ (3)</td>
<td>8.40</td>
<td>90.53</td>
<td>1326.80</td>
<td>6338+203</td>
<td>87.88</td>
<td>5.19</td>
<td>6.93</td>
<td>0.069135±0.000346</td>
<td>0.56286±0.00295</td>
<td>0.059048±0.00094</td>
<td>430.9</td>
<td>453.4</td>
<td>568.9</td>
<td></td>
</tr>
<tr>
<td>18/ 45-75μ (3)</td>
<td>5.50</td>
<td>107.51</td>
<td>1584.15</td>
<td>2839+41</td>
<td>87.18</td>
<td>5.13</td>
<td>7.69</td>
<td>0.067398±0.000338</td>
<td>0.54652±0.00285</td>
<td>0.058810±0.00088</td>
<td>420.5</td>
<td>442.7</td>
<td>560.1</td>
<td></td>
</tr>
<tr>
<td>70/ MONAZITE</td>
<td>0.71</td>
<td>1229.07</td>
<td>2284.66</td>
<td>2958±51</td>
<td>11.36</td>
<td>0.63</td>
<td>88.01</td>
<td>0.070372±0.000353</td>
<td>0.53600±0.00283</td>
<td>0.055241±0.00095</td>
<td>438.4</td>
<td>435.8</td>
<td>422.0</td>
<td></td>
</tr>
<tr>
<td>75/ MONAZITE</td>
<td>0.83</td>
<td>1324.32</td>
<td>2509.18</td>
<td>2850±27</td>
<td>11.65</td>
<td>0.65</td>
<td>87.70</td>
<td>0.070824±0.000355</td>
<td>0.54464±0.00280</td>
<td>0.055774±0.00067</td>
<td>441.1</td>
<td>441.5</td>
<td>443.4</td>
<td></td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.*
Fig. 6.9 A concordia plot of the uranium-lead data from the Numbla Vale Adamellite (Table 6.5). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data and intersects the concordia at 393.3 $^{+12.8}_{-20.6}$ m.y. and 1631 $^{+186}_{-208}$ m.y. respectively.
560.1 ± 3.3 m.y. to 781.5 ± 2.5 m.y. The data are not fitted to a straight line to within analytical error (M.S.W.D. = 4.7). No increase in the excess scatter with increasing $^{206}\text{Pb}/^{238}\text{U}$ could be detected so Model 3 is considered to be the most appropriate. The discordia calculated using this model intersects the concordia at 393.3 $^{+12.8}_{-20.6}$ m.y. and 1631 $^{+186}_{-208}$ m.y. respectively (Fig. 6.9, Line A). The age calculated from the lower intercept is significantly different from the monazite age at the 1% level and from the mean biotite age at the 5% level.

6.4 THE DALGETY GRANODIORITE

6.4.1 The granitoid

The Dalgety Granodiorite forms a large elliptical pluton (290 km$^2$ in outcrop area) in the central region of the Berridale Batholith (Fig. F.1). It has been described in detail by White et al. (1977, 1978). Because the mineralogy of the Dalgety is variable over a narrow range and approximately 65% of the total feldspar is plagioclase, the pluton consists of both adamellite and granodiorite. When the Dalgety was first mapped, it was called the Dalgety Adamellite (Lambert & White, 1965) but subsequent work has revealed an overall predominance of granodiorite so the name has been altered to the present one.

The texture of the Dalgety is unique among the S-type granitoids of the Berridale Batholith. The rock is coarse-grained and even-textured, with no discernible foliation. Characteristically, the biotite is present as large (5 mm) hexagonal grains, and pseudomorphs after cordierite are very rare. In thin section, the plagioclase is seen to occur as well formed rectangular crystals with rounded calcic cores and multiply-zoned rims as in the other S-type granitoids. Potassium feldspar occurs interstitially and is variable in abundance.

The granitoid contains relatively few xenoliths, and those that are present are of the fine-grained, even-textured type. The xenoliths have a
chemical composition which is closely related to that of the granitoid and their mineralogy is the same, but texturally they are distinctive. Fine (0.5 mm) euhedral biotite and zoned plagioclase grains are partially or completely encased in potassium feldspar and relatively large (up to 5 mm) quartz grains. This gives the xenoliths the appearance of being porphyritic in quartz. White et al. (1978) interpret this fabric as being indicative of an igneous origin and consider the xenoliths to be early crystallisation products of the granitoid which have been re-incorporated into the magma. They also note, however, that the xenoliths are low in both silica and calcium relative to the granitoid, which would not be expected if such were their origin.

The chemistry of the Dalgety Granodiorite is slightly, but significantly, different from that of the other S-type granitoids. The granitoid has a $K_2O/Na_2O$ ratio which is high relative to that in I-type granitoids, which is consistent with its having a sedimentary source, but also, $Na_2O$ and $CaO$ are high with respect to $Al_2O_3$. As a consequence, $Al/(Na + K + Ca/2)$ is lower in the Dalgety than in any other S-type granitoid in the region. It is low enough, in general, to prevent the crystallisation of cordierite but is still too high to permit the crystallisation of hornblende, either in the granitoid itself or in its xenoliths. White et al. (1978) considered these chemical features to be sufficiently characteristic of the Dalgety that its derivation from a source of the same composition as that of the other S-type granitoids was very unlikely. It was their opinion that the source of the Dalgety was unquestionably sedimentary, and that it was a very large volume of rock that had a very uniform chemical composition. A sedimentary source which was less mature than that of the other S-type granitoids was proposed.

6.4.2 The zircon

The zircon sample of the Dalgety (75-113) was collected from the same locality as geochemical sample BB9 (Fig. F.1). The zircon in the
Dalgety is very unevenly distributed between the principal minerals (Fig. 6.10). The highest concentration of zircon and the largest zircon grains occur in biotite, lesser amounts occur in plagioclase and quartz, and zircon is rare in potassium feldspar.

The zircons themselves are very similar to those in the other S-type granitoids. The majority of the grains are prismatic, with length to width ratios in the range 1:1 to 4:1, and with multifaceted terminations (Plate X(a), (b)). In many cases, the crystal faces are well developed and there is no evidence of chemical corrosion (Plate X(e), Plate XI(c), (d)) but exceptions are common (Plate X(d), (f), Plate XI(a), (b), (f)). Neither cores nor inclusions are abundant, but many examples of each were found (Plate X(c), (d)). Very few examples of parallel growth were observed (Fig. 6.10). The crystal shape of some grains suggested that their growth had been constricted (Plate X(d)).

6.4.3 Geochronology

6.4.3.1 Rubidium-strontium and potassium-argon analyses

Twenty-eight samples from the Dalgety, seven of which were xenoliths, have been analysed for total-rock rubidium-strontium geochronology. The scatter of the data about an isochron was considerably less than that observed in the Cootralantra but it was similarly beyond the limits of experimental error (M.S.W.D. = 31.2). The scatter was not reduced significantly when the xenolith data were excluded from the regression (M.S.W.D. = 29.7), which supports the proposal of White et al. (1978) that the xenoliths are early crystallisation products of the magma. The regression analysis of the excess scatter showed that it increased as a function of $\frac{^{87}\text{Rb}}{^{86}\text{Sr}}$ and hence that Model 2 was the most appropriate. The age calculated from the combined xenolith and granitoid total-rock data using this model is $426.8 \pm 11.2$ (t s.e.) m.y.

Six of the freshest samples of the Dalgety were selected for biotite rubidium-strontium age determinations, and four of those were dated
Fig. 6.10 The distribution of zircon in the minerals of the Dalgety Granodiorite, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
FIGURE 6.10

DALGETY GRANODIORITE

K-FELDSPAR

BIOTITE

PLAGIOCLASE

QUARTZ

100 \mu
Plate X  Light photomicrographs of zircons from the Dalgety Granodiorite.

Scale subdivisions are 10μ. The fractions illustrated are:

(a) 45 – 75μ (4)  (b) 45 – 75μ (3)
(c) +100μ (4')  (d) 75 -100μ (4)
(e) +100μ (4')  (f) 75 -100μ (4)
Plate XI Scanning electron photomicrographs of zircons from the Dalgety Granodiorite. Scale subdivisions are 30μ. The zircons are from fraction 45 – 75μ (3).
by potassium-argon. By neither method was any significant difference in biotite age between the samples detected. The mean biotite age measured by rubidium-strontium was $416.7 \pm 1.3$ m.y. and measured by potassium-argon was $414.8 \pm 3.3$ m.y. The difference between these ages is not significant at the 5% level. The weighted mean biotite age of the Dalgety is $416.4 \pm 1.2$ m.y. This is not significantly different from the total-rock age at the 5% level.

6.4.3.2 Uranium-lead analyses

Both zircon and monazite from the Dalgety were analysed for uranium-lead. The data are listed in Table 6.6 and plotted on a concordia diagram in Fig. 6.11. The more precise of the duplicate monazite analyses (76/M) plots significantly above the concordia, and when the apparent ages from the mean values of $\frac{207}{206}$Pb/$\frac{206}{206}$Pb, $\frac{206}{238}$Pb/$\frac{238}{238}$U and $\frac{207}{235}$Pb/$\frac{235}{235}$U for the duplicate are calculated, the first ($392.0^{+22.4}_{-22.7}$ m.y.) is found to be significantly different from the latter two ($429.8 \pm 3.0$ m.y. and $424.1 \pm 4.5$ m.y. respectively) at the 5% level. The discordance cannot be explained by uncertainty in the uranium decay constants (the uncertainty in a $\frac{207}{206}$Pb/$\frac{206}{206}$Pb age of 392 m.y. was calculated by Stern (1972) to be 2.9 m.y.).

If the discordance is considered to be due to a recent disturbance of the monazite's lead to uranium ratio, the $\frac{207}{206}$Pb/$\frac{206}{206}$Pb age of $392.0^{+22.4}_{-22.7}$ m.y. would be the best estimate of the monazite age. However, because the duplicate determinations of $\frac{206}{238}$Pb/$\frac{238}{238}$U and $\frac{207}{235}$Pb/$\frac{235}{235}$U were the same to within analytical error respectively, such a disturbance is thought to be unlikely.

As with the monazite analyses from the Numbla Vale, it is considered that the error in the determination of $\frac{207}{207}$Pb has been underestimated and that the best estimate of the monazite age is that calculated from $\frac{206}{238}$Pb/$\frac{238}{238}$U, $429.8 \pm 3.0$ m.y. This age is significantly older than the Dalgety's biotite age ($416.4 \pm 1.2$ m.y.) at the 1% level.

The pattern of discordance for zircon analyses from the Dalgety differs from the pattern obtained from the other S-types. The effects of
Table 6.6 Uranium-lead analytical data from the Dalgety Granodiorite (75-113). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar from the Numbla Vale Adamellite (Appendix E).
**TABLE 6.6**

**ANALYTICAL DATA FOR THE DALGETY GRANODIORITE (75-113)**

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight mg</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206 Pb* measured</th>
<th>206 Pb</th>
<th>207 Pb</th>
<th>208 Pb</th>
<th>206 Pb</th>
<th>207 Pb</th>
<th>208 Pb</th>
<th>238 U</th>
<th>235 U</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>52/ +100µ (4')</td>
<td>2.33</td>
<td>30.51</td>
<td>389.66</td>
<td>5444+110</td>
<td>85.67</td>
<td>5.86</td>
<td>8.47</td>
<td>.077670+ .000388</td>
<td>.73238+ .00367</td>
<td>.068388+ .000024</td>
<td>482.2</td>
<td>557.9</td>
<td>880.3</td>
</tr>
<tr>
<td>55/ 45-75µ (4)</td>
<td>3.10</td>
<td>32.13</td>
<td>456.13</td>
<td>5374+173</td>
<td>85.88</td>
<td>5.39</td>
<td>8.73</td>
<td>.070061+ .000350</td>
<td>.60606+ .00317</td>
<td>.062739+ .000097</td>
<td>436.5</td>
<td>481.1</td>
<td>699.4</td>
</tr>
<tr>
<td>53/ 75-100µ (4)</td>
<td>9.31</td>
<td>36.65</td>
<td>502.55</td>
<td>9614+324</td>
<td>86.03</td>
<td>5.59</td>
<td>8.38</td>
<td>.072771+ .000365</td>
<td>.65150+ .00333</td>
<td>.064931+ .000071</td>
<td>452.8</td>
<td>509.4</td>
<td>772.1</td>
</tr>
<tr>
<td>51/ +100µ (4&quot;)</td>
<td>10.68</td>
<td>40.57</td>
<td>544.53</td>
<td>10513+115</td>
<td>86.21</td>
<td>5.79</td>
<td>8.00</td>
<td>.074419+ .000372</td>
<td>.68978+ .00346</td>
<td>.067224+ .000020</td>
<td>462.7</td>
<td>532.7</td>
<td>844.7</td>
</tr>
<tr>
<td>92/ 45-75µ (3)</td>
<td>1.36</td>
<td>39.94</td>
<td>679.69</td>
<td>3728+59</td>
<td>86.49</td>
<td>5.14</td>
<td>8.37</td>
<td>.058797+ .000296</td>
<td>.48159+ .00253</td>
<td>.059405+ .000102</td>
<td>368.3</td>
<td>399.2</td>
<td>582.0</td>
</tr>
<tr>
<td>56/ -45µ (3)</td>
<td>1.87</td>
<td>49.13</td>
<td>864.78</td>
<td>2139+39</td>
<td>85.53</td>
<td>4.96</td>
<td>9.51</td>
<td>.055828+ .000279</td>
<td>.44626+ .00246</td>
<td>.057975+ .000136</td>
<td>350.2</td>
<td>374.7</td>
<td>528.8</td>
</tr>
<tr>
<td>54/ 45-75µ (3)</td>
<td>1.78</td>
<td>51.92</td>
<td>961.31</td>
<td>3051+75</td>
<td>86.47</td>
<td>5.13</td>
<td>8.40</td>
<td>.053931+ .000270</td>
<td>.44096+ .00245</td>
<td>.059301+ .000143</td>
<td>338.6</td>
<td>370.9</td>
<td>578.2</td>
</tr>
<tr>
<td>71/ MONAZITE</td>
<td>1.32</td>
<td>998.65</td>
<td>724.15</td>
<td>1285+54</td>
<td>4.35</td>
<td>0.24</td>
<td>95.41</td>
<td>.068998+ .000348</td>
<td>.52297+ .00616</td>
<td>.054972+ .000588</td>
<td>430.1</td>
<td>427.1</td>
<td>411.1</td>
</tr>
<tr>
<td>76/ MONAZITE</td>
<td>1.04</td>
<td>1009.38</td>
<td>745.97</td>
<td>1779+47</td>
<td>4.42</td>
<td>0.24</td>
<td>95.34</td>
<td>.068894+ .000345</td>
<td>.51652+ .00392</td>
<td>.054376+ .000310</td>
<td>429.5</td>
<td>422.8</td>
<td>386.6</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.
Fig. 6.11 A concordia plot of the uranium-lead data from the Dalgety Granodiorite (Table 6.6). The estimates of analytical error are two standard errors. Line A is the line of best fit to the data for fractions 51/, 52/, 53/ and 55/ and intersects the concordia at 378.8 ±21.2 m.y. and 1751 ±269 m.y. and 1751 ±269 m.y. respectively.
FIGURE 6.11

DALGETY
GRANODIORITE
(75-113)
inheritance are strong, but rather than approximating to a straight line, the data appear to define a curve (Fig. 6.11). The zircon fractions with relatively low concentrations of uranium ($51/_{\text{th}}, 52/_{\text{th}}, 53/_{\text{th}}$ and $55/_{\text{th}}$) are almost aligned, although they are not in fact fitted to a straight line to within analytical error. The intercepts of the line of best fit (Fig. 6.11, Line A) and the concordia are equivalent to ages of $378.8^{+21.2}_{-41.6}$ m.y. and $1751^{+269}_{-317}$ m.y. respectively. The fractions with higher uranium concentrations ($54/_{\text{th}}, 65/_{\text{th}}$ and $92/_{\text{th}}$) have much lower $^{207}\text{Pb}/^{206}\text{Pb}$ ratios and plot well below this line. $378.8^{+21.2}_{-41.6}$ m.y. is significantly less than both the monazite and biotite ages at the 5% level.

6.5 THE BOG HOLE ADAMELLITE

6.5.1 The granitoid

The Bog Hole Adamellite is a relatively small intrusive, with an outcrop area of $10 \text{ km}^2$, which is located near the northern end of the Kosciusko Batholith (Fig. F.1). It has been described in detail by White et al. (1977). The Bog Hole is one of the most felsic S-type granitoids in the Snowy Mountains region, and in places is not adamellite but granite sensu stricto. In hand specimen, the granitoid appears to be coarse-grained. The average grainsize is about 5 mm, but the potassium feldspars commonly are larger than 1 cm. Although the rock is very felsic, pinite pseudomorphs after cordierite are common. In thin section, the grainsize is seen to be in fact much smaller. The large quartz grains are aggregates of smaller grains which are of the order of 2 mm in diameter. Undulose extinction in the quartz is strong. The large potassium feldspars commonly are fractured and the fractures are filled in by quartz. These features suggest that the pluton has been under stress at some time since it crystallised. No foliation is developed, however. The plagioclase grains generally are weakly zoned and have strongly altered cores. The biotite also commonly is altered and is to a large extent replaced by chlorite and epidote.

Because the Bog Hole is part of the Kosciusko Batholith, it was not included in the total-rock rubidium-strontium study of Compston et al.
(1977a) or in the biotite rubidium-strontium study of Williams et al. (1976). As a consequence, the granitoid's geochronology is very poorly known. It was selected for zircon analysis because it was the most felsic S-type granitoid in the region and was considered, therefore, to contain a minimal amount of solid residual material from its source. The aim of the zircon study was to determine whether this would be reflected in the amount of inheritance observed.

6.5.2 The zircon

The zircon sample from the Bog Hole (77-040) was the most felsic which could be obtained from the pluton and was a granite sensu stricto. It was collected from the same locality as geochemical sample KB42 (Fig. F.1). In thin section, the zircon is seen to be concentrated strongly in the biotite and quartz (Fig. 6.12). Lesser amounts were found in plagioclase and very little in potassium feldspar.

The zircon crystals in the Bog Hole morphologically are very similar to the zircon in the other S-type granitoids and different from the zircons from the very felsic I-type granitoids (the Maffra and the Delegate). The zircons are prismatic, with length to width ratios in the range 1:1 to 4:1 and have multifaceted terminations (Plate XII). Some grains are well-formed crystals (Plate XII(b), (f)) but the majority show the effects of chemical corrosion (Plate XII(a), (c), (d)), more so than do the zircons from the other S-type granitoids. Rare examples of cores, inclusions and parallel growth were found (Plate XII(c), (e)).

6.5.3 Geochronology

6.5.3.1 Uranium-lead analyses

The only geochronological technique which has been applied to the Bog Hole is zircon uranium-lead; the granitoid is not monazite-bearing. The analytical data are listed in Table 6.7 and are plotted in Fig. 6.13. The data are strongly discordant and the range of measured \(^{207}\text{Pb}/^{206}\text{Pb}\) ages is
Fig. 6.12 The distribution of zircon in the minerals of the Bog Hole Granite, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
FIGURE 6.12

K-FELDSPAR

BIOTITE

PINITE

PLAGIOCLASE

BOG HOLE

GRANITE

QUARTZ

100μ
Plate XII  Light photomicrographs of zircons from the Bog Hole Granite.

Scale subdivisions are 10μ. The fractions illustrated are:

(a)  60 - 100μ (1)  (b)  60 - 100μ (4)
(c)  - 60μ (2)      (d)  60 - 100μ (3)
(e)  60 - 100μ (3)  (f)  + 100μ (2)
Table 6.7 Uranium-lead analytical data from the Bog Hole Granite (77-040). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar from the Numbla Vale Adamellite (Appendix E).
### Table 6.7

**Analytical Data for the Bog Hole Granite (77-040)**

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight (mg)</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>$^{206}\text{Pb}$/$^{204}\text{Pb}$ Measured</th>
<th>Atom Percent Radiogenic Lead</th>
<th><em>Atomic Ratios</em></th>
<th>Apparent Ages (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>206\text{Pb}</td>
</tr>
<tr>
<td>101/ +100μ (2)</td>
<td>0.95</td>
<td>31.45</td>
<td>375.28</td>
<td>3092±527</td>
<td>85.64</td>
<td>5.57</td>
<td>8.79</td>
</tr>
<tr>
<td>100/ 60 -100μ (4)</td>
<td>1.69</td>
<td>42.82</td>
<td>479.94</td>
<td>10892±1986</td>
<td>86.00</td>
<td>6.26</td>
<td>7.74</td>
</tr>
<tr>
<td>98/ 60 -100μ (3)</td>
<td>1.28</td>
<td>39.94</td>
<td>510.62</td>
<td>4181±179</td>
<td>86.78</td>
<td>5.76</td>
<td>7.46</td>
</tr>
<tr>
<td>97/ -60μ (2)</td>
<td>2.94</td>
<td>69.42</td>
<td>994.38</td>
<td>9906±71</td>
<td>89.41</td>
<td>5.28</td>
<td>5.31</td>
</tr>
<tr>
<td>99/ 60 -100μ (1)</td>
<td>1.44</td>
<td>69.30</td>
<td>1006.18</td>
<td>7946±26</td>
<td>88.54</td>
<td>5.09</td>
<td>6.37</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.*
Fig. 6.13  A concordia plot of the uranium-lead data from the Bog Hole Granite (Table 6.7). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data and intersects the concordia at 428.7 $^{+4.3}_{-5.0}$ m.y. and 1926 $^{+76}_{-80}$ m.y. respectively.
larger than that observed in any other S-type granitoid, $513.0 \pm 8.8$ m.y. (99/) to $1008.1 \pm 8.1$ m.y. (100/). The points are not fitted to a straight line to within analytical error ($M.S.W.D. = 5.3$) and an analysis of the excess scatter showed that it increased significantly with increasing $^{206}\text{Pb}/^{238}\text{U}$; Model 2 is therefore the most appropriate. The discordia calculated using Model 2 (Fig. 6.13, Line A) intersects the concordia at $428.7^{+4.3}_{-5.0}$ m.y. and at $1926^{+76}_{-80}$ m.y.

6.6 INTERPRETATION OF THE GEOCHRONOLOGY

The principal problem in the interpretation of the geochronology of the S-type granitoids is to determine which of the measured ages are significant geologically and, for those that are, to determine what that significance is. Intrusive relationships and stratigraphic constraints can be used to place limits on the plutons' emplacement ages and so to reduce the number of possible interpretations. No similar limit on possible source ages is available however, and whether or not the measured ages are in fact source ages must be deduced from the isotopic data alone.

The most important single feature of the zircon discordance patterns from the S-type granitoids is the wide range of $^{207}\text{Pb}/^{206}\text{Pb}$ ages which is obtained from each. If such discordance is interpreted using the episodic lead loss model first proposed by Wetherill (1956a,b), the age indicated by the upper intercept of the discordia and concordia is the age of primary crystallisation, and the age indicated by the lower intercept is the time at which an episode of lead loss occurred. The range of upper intercept ages determined on the S-type granitoids is from $1343^{+213}_{-228}$ m.y. for the Cootralantra to $1926^{+76}_{-80}$ m.y. for the Bog Hole. One possible interpretation is that these ages are emplacement ages and that the ages indicated by the lower intercepts, all of which are approximately 400 m.y., are ages of episodic disturbance. The age of the Batholith's country rock can be used to determine whether such emplacement ages are possible.

From intrusive relationships, the Cootralantra is one of the
oldest S-type granitoids in the Berridale Batholith. It is intrusive into
the Adaminaby Beds and graptolites have been found within the pluton's
contact aureole (Chappell & White, 1976). An older limit to the emplacement
age of the Cootralantra is Eastonian, which in absolute terms is probably
460 ± 10 m.y. (see section 4.3.4). Near the northern margin of the
Cootralantra, the Adaminaby Beds contain an infolded unit of felsic
volcanics (the Bennetts Creek Volcanics) which are demonstrably stratigraphi-
ically higher than the surrounding Ordovician rocks (Hopwood, 1969). White
et al. (1977) consider that this unit has been refolded during the emplace-
ment of the Cootralantra. If this is so, and if the volcanics are Late
Llandoveryian as White et al. (1977) tentatively suggest, the lower limit to
the emplacement age of the Cootralantra is in fact 437 m.y. (Lanphere et al.'s
(1977) estimate for the age of the base of the Silurian). In the light of
these facts, the discordia upper intercept ages (1343 to 1926 m.y.) cannot
be the plutons' ages of emplacement. The zircon discordance must be due, in
large part, to inheritance.

If the discordance is interpreted as an inheritance phenomenon,
one possible model is that it is generated principally by the mixing of two
components: zircon crystallised from the granitoid magma and zircon which
has been inherited. If the latter component is homogeneous, a single mixing
line would result, but if it is itself a mixture of zircon with a variety
of primary ages, the discordance is best modelled by a series of mixing lines
which intersect at the primary age of the magmatic zircon. (A model similar
to this was used by Pidgeon (1969) to explain the zircon discordance in
sediments from the Dalradian of Ireland).

The significant feature of the above model is that the age of the
discordia's lower intercepts is the age of the magmatic zircon. It would
be expected, therefore, that this age would be an older limit to the
pluton's emplacement age. However, in each of the plutons for which monazite
and/or biotite ages are available, the ages of the lower intercept is equal
to, or significantly younger than those ages. Because of the very low
blocking temperature of biotite for both rubidium-strontium and potassium-argon (possibly as low as 300°C (Jäger & Zwart, 1968)) it is extremely unlikely that the closure of the biotite systems has predated magmatic zircon crystallisation; either the zircon lower intersection ages record a later, very low-temperature event or those ages are not significant geologically. The latter is considered the more likely for two reasons: first, it can be argued from the Cootralantra data that there was not a single younger event, and secondly, the discordance pattern from the Dalgety puts the validity of the model itself in doubt.

In the discussion of the Cootralantra granitoid above (section 6.1.3.2), it was noted that the ten analyses of zircon fractions were not fitted to a straight line to within analytical error (Fig. 6.2). This is the observation which would be predicted if the discordance were due to older zircons of different ages being strongly disturbed by a younger event. However, contrary to that model, the analysis of the excess scatter detected no significant increase in the scatter as a function of increasing $^{206}\text{Pb}/^{238}\text{U}$. This does not preclude disturbance of the zircon by a single event but it does argue strongly against it.

The non-linear discordance pattern for the Dalgety zircon (Fig. 6.11) suggests a likely explanation for the young lower intersection ages. The discordance in the Dalgety closely resembles the discordance pattern generated by the combined data from the central Swiss Alps, compiled by Allègre et al. (1974), and the curve defined by the Dalgety data is very similar to the curves which those authors calculated on the basis of multi-episodic lead loss models. An important implication of Allègre et al.'s (1974) calculations which is relevant to the interpretation of the S-type granitoids' zircon data is that in cases where multiepisodic lead loss has occurred, the extrapolation of the discordia to a lower intersection with the concordia does not necessarily indicate the age of any particular loss episode.

As far as is known, the Berridale Batholith has not been subjected
to repeated metamorphism so a multiepisodic model cannot be applied to the interpretation of the zircon data directly. Nevertheless, the discordance pattern for the Dalgety shows that in that pluton at least, some disturbance of the zircon must be considered. In view of the evidence in the discordance patterns of the Tara, the Maffra and the Delegate for a recent episode of net lead loss, recent lead loss is considered to be the most likely explanation.

Because no zircon fraction analysed from the S-type granitoids can be assumed to be free of the effects of lead loss, and in the light of the arguments above, the ages indicated by the lower intersections of the concordia and the S-type granitoids' discordias are considered to be minimum estimates of the age of the younger zircon component. Since it is not possible to estimate the extent of the recent lead loss, for the purposes of interpreting the geochronology, these ages are not significant geologically. This conclusion puts the geological significance of the age of the upper intersection in doubt.

Zircon discordance patterns due to inheritance have been reported by a number of authors and the significance of the upper intersection age has been variously interpreted. When the pattern was first observed (Grauert & Arnold, 1968), the upper intercept was interpreted as the age of a detrital component. Some later authors were more cautious in attributing the age geological significance (Pidgeon, 1969; Pidgeon et al., 1970) but as more data has become available, consistencies in the upper intersection ages from particular regions, notably the Swiss Alps, have become apparent (Pidgeon et al., 1970; Köppel & Grünenfelder, 1971; Grauert et al., 1973; Allègre et al., 1974; Köppel, 1974 and Gulson & Rutishauser, 1976), and these ages have come to be accepted as the primary age of the zircon from the rocks' source.

There is not sufficient data from the Snowy Mountains region for the constancy of upper intersection ages to be used as an argument in favour of their being significant geologically, although the ages which have been
determined fall within a relatively narrow range (1343±213 to 1926±76 m.y.). For this reason, and because it is thought to be unlikely that the mixing of magmatic zircon and discordant detrital zircon of several ages, followed by a variable net loss of lead, would produce a mixing line which, when extrapolated to the concordia, indicates the age of the older components, the ages of the upper intercepts of the S-type granitoids' discordias are not considered necessarily to be the age of the inherited component's provenance.

If it is assumed that the zircon discordance pattern in the S-type granitoids is due to the mixing of inherited zircon and zircon crystallised from the granitoid magma, followed by recent lead loss, the oldest 207Pb/206Pb age measured on zircon from each granitoid is a younger limit to the 207Pb/206Pb age of the oldest component in the mixture. By this argument, the provenance of the oldest component must be equal to or older than 893.8±3.0 m.y. for the Cootralantra, 1163.1±4.5 m.y. for the Cootralantra xenolith, 997.0±25.3 m.y. for the Bullenbalong, 781.5±2.5 m.y. for the Numbla Vale, 880.3±1.5 m.y. for the Dalgety and 1008.1±8.1 m.y. for the Bog Hole. Even these ages are considerably older than the age of any of the rocks presently exposed in the Lachlan Mobile Zone. The question which this raises is what is the source of the older component.

One possible explanation of the older component in the S-type granitoids is that it has been contributed by country rock which has been assimilated by the magma. This is considered to be unlikely, however. First, if the amount of inheritance required to produce the observed discordance is calculated (for fraction 79/, for example), assuming the magmatic zircon to be 420 m.y. old (the biotite age of the Cootralantra), this amount is found to be very large (Fig. 6.14). Even if the 207Pb/206Pb age of the inherited component is 2000 m.y., more than 30% of the 207Pb observed in the zircon must be contributed by inheritance. The absence of evidence for the assimilation of any country rock by the I-type granitoids argues very strongly against it being possible to explain this inheritance by assimilation alone. Secondly, the chemical composition of the S-types indicates that they are derived from metasediments and these would be
Fig. 6.14 The relationship between the proportion of the total $^{207}\text{Pb}$ which is contributed by inheritance and the $^{207}\text{Pb}/^{206}\text{Pb}$ age of the inherited component for fraction 79/. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of the lead in the magmatic zircon is assumed to be 420 m.y.
LEAD INHERITANCE

in

FRACTION 79/

FIGURE 6.14

207\text{Pb}/206\text{Pb} age of inherited component (b.y.)

Inherited 207\text{Pb} (%)
expected to contain older, detrital zircon. Thirdly, the zircon analysed from the Cootralantra xenolith (which is considered to be a piece of the source material) showed equally strong inheritance to the granitoid. This is compatible with the observation that the amount of inheritance observed is independent of the amount of solid residual source material that each granitoid contains. There is therefore no need to propose the assimilation of country rock to explain the discordance patterns observed.

In the light of the above arguments, it is a prerequisite of any postulated source for the S-type granitoids that it contain zircon with \( {^{207}\text{Pb}/^{206}\text{Pb}} \) ages at least as old as the oldest found in the granitoids, namely \( 1163.1 \pm 4.5 \) m.y. To test the proposal that the local Ordovician sediments were the granitoids' source, zircon was extracted from a sample of the Batholith's country rock and analysed for uranium-lead.

6.6.1 The Dalgety Sandstone

A sample of sandstone (76-368) was collected from the sedimentary screen which separates the Berridale from the Kosciusko Batholith (Fig. F.1). It was collected from the Adaminaby Beds at a locality where the sedimentary section was well known. The sandstone was underlain by black slate containing an assemblage of Eastonian graptolites, and the sequence which overlay the sandstone included a second unit of black slate in which Bolindian graptolites had been found (White et al., 1977). There is no doubt that the sandstone was deposited in the Late Ordovician.

In thin section, the sandstone was seen to be partially recrystallised; the interstices between the grains were filled by quartz but the rounded margins of the original quartz grains were clearly defined by rings of impurities. The zircon in all cases was found in the interstices.

The zircons themselves were very strongly rounded and had a characteristic surface texture which is considered to be the result of abrasion (Plates XIV, XV). The length to width ratios of the grains were very similar to those observed in zircon from the S-type granitoids, 1:1 to
3:1, but in contrast to the latter, on very few grains were crystal faces observed (Plates XIV(a), (b)). An indication of the possible complexity of the history of the zircon is the presence of cores of even older zircon in some of the grains (Plate XIV(f), Plate XV(a)).

The zircon population was a mixture of a variety of zircon types, and in preparing fractions for analysis, an attempt was made to isolate zircon which might have a common provenance. The criterion chosen for the separation was colour (Appendix G). The uranium-lead analyses of zircon from the Dalgety Sandstone are listed in Table 6.8 and plotted on a concordia diagram in Fig. 6.15. The data are strongly discordant and the measured \( \frac{^{207}Pb}{^{206}Pb} \) ages (1442 to 1823 m.y.) all exceed the maximum \( \frac{^{207}Pb}{^{206}Pb} \) age of zircon in the granitoids (1163 m.y.). It is concluded therefore, that the age of the zircons in the Adaminaby Beds is compatible with the proposal that those rocks are the source of the S-type granitoids.

A second prerequisite of any source proposed for the S-type granitoids is that it must be possible chemically to derive them from it. A study of the chemical composition of sedimentary rocks in the Snowy Mountains region which is currently in progress has revealed consistent chemical differences between those rocks and the S-type granitoids which make it extremely unlikely that they are the S-types' source, even allowing for the possibility that other rocks such as carbonates and mafic volcanics may also have been melted (L. Wyborn pers. comm., 1977). In view of these findings, the chemistry of the S-type granitoids, the zircon data obtained from the S-type granitoids in the present study and the abrupt transition from S-type to I-type granitoids across the I-S line (White et al., 1976), the evidence for metasedimentary basement underlying the Snowy Mountains region is considered to be very strong. The zircon data do not permit a limit to be placed on the age of that basement but they do indicate that its provenance must be at least as old as 1163 m.y. This age is similar to the minimum provenance age proposed for sediments of the Greenland Series, New Zealand by Aronson (1968) on the basis of zircon \( \frac{^{207}Pb}{^{206}Pb} \) ages.
Table 6.8 Uranium-lead analytical data from the Dalgety Sandstone (76-368). The estimates of precision were calculated as outlined in Appendix C. The common lead composition was calculated for an age of 415 m.y. using the 'linear' Model II of Cumming and Richards (1975).
## TABLE 6.8

ANALYTICAL DATA FOR THE DALGETY SANDSTONE (76-368)

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight (mg)</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>ATOM PERCENT RADIOGENIC LEAD</th>
<th>ATOMIC RATIOS</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>206(^{\text{Pb}})(^{\text{measured}})</td>
<td>206(^{\text{Pb}}) / 238(^{\text{U}})</td>
<td>207(^{\text{Pb}}) / 235(^{\text{U}})</td>
</tr>
<tr>
<td>50/ +100(\mu) Mixed</td>
<td>2.26</td>
<td>65.52</td>
<td>562.47</td>
<td>2719±11</td>
<td>80.88</td>
<td>8.32</td>
</tr>
<tr>
<td>47/ 75-100(\mu) Pink</td>
<td>3.75</td>
<td>62.68</td>
<td>584.95</td>
<td>2754±51</td>
<td>81.62</td>
<td>7.41</td>
</tr>
<tr>
<td>49/ 60-75(\mu) Purple</td>
<td>0.98</td>
<td>96.21</td>
<td>627.48</td>
<td>1870±71</td>
<td>81.49</td>
<td>9.08</td>
</tr>
<tr>
<td>48/ 60-75(\mu) Pink</td>
<td>2.46</td>
<td>66.16</td>
<td>640.63</td>
<td>2807±73</td>
<td>80.95</td>
<td>7.55</td>
</tr>
<tr>
<td>46/ -60(\mu)</td>
<td>10.02</td>
<td>55.71</td>
<td>663.31</td>
<td>4039±250</td>
<td>81.40</td>
<td>7.81</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.*
Fig. 6.15 A concordia plot of the uranium-lead data from the Dalgety Sandstone (Table 6.8). The estimates of analytical error are two standard errors. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of fraction 49/ (1823 ± 11 m.y.) is a younger limit to the age of the oldest rock from which the Sandstone was derived.
FIGURE 6.15

SANDSTONE
(76-368)

$^{207}Pb / ^{235}U$ vs. $\nu$
Because of the complexity of the zircon discordance patterns and the impossibility of correcting the data for the effects of recent lead loss, the zircon analyses provide no more than younger limits to the S-types' emplacement ages. The problem of dating emplacement must be solved using data from monazite and biotite.

Williams et al. (1976) argued strongly on consistency grounds that the rubidium–strontium biotite ages from the Berridale Batholith were emplacement ages. They pointed out that within any one pluton, the biotite ages were identical to within analytical error, but that significant differences between the ages of different plutons could be detected. (The nested classification analysis of variance on which these conclusions were based is described in Appendix D and repeated with the inclusion of additional, more recent analyses). Where the ages were different, they were without exceptions compatible with the intrusive sequence which had been established geologically. The same appears to be true of biotite potassium–argon ages, although that work has yet to be completed (N.W. Tetley pers. comm., 1977).

A similar test of consistency can be applied to the monazite ages. The relative ages of three of the S-type plutons for which monazite ages have been measured may be deduced from intrusive relationships. The Numbla Vale is older than the Dalgety, which in turn both intrudes and is intruded by the Cootralantra (Fig. 2.1). The monazite ages are consistent with this sequence; the monazite ages of the Cootralantra and the Dalgety are not detectably different from one another (427.9 ± 3.0 m.y. and 429.8 ± 3.0 m.y. respectively), but both are significantly younger than the monazite age of the Numbla Vale (439.8 ± 3.0 m.y.) at the 5% level.

Although both the monazite and biotite ages of the plutons are consistent with the intrusive sequence and, within the stratigraphic constraints, both could be interpreted as recording the emplacement age, for every pluton for which both monazite and biotite ages have been determined, the latter is significantly younger than the former at the 1% level. There are three possible explanations of this observation which must be considered:
first, the monazites may record the ages of emplacement and the biotites may have been reset, secondly, the monazites may record the ages of emplacement and the plutons may have cooled slowly to the biotite blocking temperature, and thirdly, the biotites may record the ages of emplacement and the monazites may have closed beforehand.

The constancy of biotite ages within plutons and the significant differences between plutons argue strongly against regional resetting (Williams et al., 1976). The possibility of slow cooling however, must be examined more closely.

Very little information is presently available on the blocking temperature of monazite for uranium and lead. In most studies, the age of monazite has been interpreted as its crystallisation age, whether it crystallised at the time its host rock was emplaced (for example, Gulson and Krogh, 1973) or at the same time of a metamorphic event (for example, Hänny et al., 1975). Köppel and Grünenfelder (1975) however, attempted to establish the effect of high temperatures on monazite ages by a study of monazites in metamorphic rocks of the Central Swiss Alps.

Köppel and Grünenfelder examined the distribution of monazite ages in the same area of the Alps that had been studied previously by Jäger et al. (1967), Jäger et al. (1969) and others, by biotite rubidium–strontium. In that earlier work, progressive changes in the biotite ages of gneisses were found, which could be related directly to changes in the metamorphic grade. Köppel and Grünenfelder found that the monazite ages in the gneisses also were not uniform, but that the variability was related, not to the samples' metamorphic grade, but to their tectonic position. The metamorphic grades which were studied were as high as amphibolite facies. In addition, a difference in monazite ages was found between a xenolith and its host granitoid. These observations were interpreted as an indication that it was possible for monazite to retain a significant proportion of its radiogenic lead, even at temperatures above 600°C.

In the light of these conclusions, it is possible that the
difference between the monazite and biotite ages in the Berridale Batholith is a consequence of slow cooling, but it is equally possible that the difference between the ages is an indication of the time interval between the crystallisation of monazite and cooling to the biotite blocking temperature. The former alternative can be tested by an examination of the differences between ages measured by other techniques.

Studies of the effects of progressive metamorphism on potassium-argon ages (for example, Hanson & Gast, 1967 and Hart et al., 1968) have shown that the blocking temperature of hornblende is much higher than that of biotite (about 500°C and 300°C respectively). It would therefore be expected that if the Berridale Batholith plutons had cooled slowly from their emplacement temperature (probably about 700°C, if the melts were water rich (Robertson & Wyllie, 1971)), through the blocking temperature of monazite (greater than 600°C), through the blocking temperature of hornblende (about 500°C) and through the blocking temperature of biotite (about 300°C), this would be reflected in a progression of ages measured on those minerals. It is found however, that although there is in every case a significant difference between the monazite and biotite ages, for the two plutons for which hornblende ages have been measured, the Tara and the Finister, the hornblende and biotite ages are not detectably different from one another. It can be argued that both the hornblende and biotite ages of the Finister have been reset by later plutons (see Chapter 7) but in the case of the Tara, the evidence against later resetting is strong (see Chapter 5). For the Tara at least, cooling from the hornblende to the biotite blocking temperature has occurred over a time interval shorter than that which can be resolved (less than 8 m.y.). There are other isotopic data which argue for cases of rapid cooling. Williams et al. (1976) reported biotite, muscovite and total-rock rubidium-strontium analyses from one pluton (the Namungo) which showed that the time interval between emplacement (as registered by the total-rock age) and cooling to the biotite blocking temperature could not exceed about 3 m.y. In addition, the biotite and monazite ages of the Maffra
are the same to within analytical error.

Although from the data which are available at present, it is not possible to prove that all the plutons in the Berridale Batholith cooled rapidly after they were emplaced, there is a strong indication that they did so. In that case, it is considered to be unlikely that the measured time intervals between the closure of monazite and the closure of biotite are a consequence of slow, post-emplacement cooling. It is concluded that the oldest biotite ages measured on individual plutons are minimum estimates of the pluton's emplacement age and that the monazite ages are ages of monazite crystallisation, which may have occurred at or before emplacement. The implication of this conclusion is that the monazite ages are minimum estimates of the age of the melting event which produced the granitoid magma.

If this model is applied to the S-type granitoids, the range of emplacement ages indicated by the biotite ages is very small (416.4 ± 1.2 to 419.0 ± 1.8 m.y. (excluding the Bullenbalong, for which later resetting is a strong possibility)) but the range of pre-emplacement, monazite crystallisation ages is large (427.9 ± 3.0 to 439.8 ± 7.0 m.y.). Similarly to the I-type magmas, it would appear that the S-type magmas have been generated over a period of time but have been emplaced almost simultaneously.

The Dalgety, the Numbla Vale and the Cootralantra are the only S-type plutons which have been analysed for uranium-lead and for which a total-rock rubidium-strontium age has been measured. The total-rock ages of the Numbla Vale and the Dalgety are not significantly different from those plutons' biotite ages at the 5% level. The total-rock age of the Cootralantra however, is significantly older than both the biotite and the monazite ages at the 1% level if it is calculated for the granitoid samples alone. It is possible to explain this difference by a separation of the Cootralantra magma from its source at 471.4 ± 21.1 m.y. followed by crystallisation and closure of the total-rock system and then remobilisation of the magma and total extraction of rubidium and strontium at 427.9 ± 3.0 m.y. but a model of
incomplete equilibration of strontium during magma genesis and inheritance of an 'isochron' from the source, similar to that proposed by Roddick (1974), is preferred. Because the extent of strontium equilibration is not known, the total-rock age does not necessarily have any significance geologically, except that it is a younger limit to the age of the granitoid's source.

6.7 SUMMARY

The feature in common to the zircon data from all the S-type granitoids is strong discordance due to a wide range in $^{207}\text{Pb}/^{206}\text{Pb}$. This is attributable to the presence of an older inherited zircon component from the granitoids' metasedimentary source rocks. The amount of this component which is present is independent of the amount of solid residual source material in the magma.

Because all the zircons in the granitoids have suffered unknown amounts of recent lead loss, the age of the discordia's lower intercept with the concordia is considered to be a minimum estimate of the age of zircon crystallised from the magma, and the age of the upper intercept is considered to be of doubtful significance geologically.

On chemical grounds it can be argued that the source of the granitoids is unlikely to be of the same composition as any of the sedimentary rocks exposed in the Snowy Mountains region, so that source is probably metasedimentary basement. From the maximum $^{207}\text{Pb}/^{206}\text{Pb}$ age measured on the zircons, the provenance of that basement has an age equal to or greater than $1163 \pm 5$ m.y.

In every S-type granitoid, the monazite age is significantly older than the biotite age (measured by potassium-argon or rubidium-strontium) at the 1% level. Because the plutons are considered to have cooled rapidly after they were emplaced and because no post-emplacement metamorphism has occurred, the biotite ages are interpreted as younger limits to the age of emplacement. The monazite ages are interpreted as monazite crystallisation ages and are younger limits to the age of the melting events which produced
the magmas. The difference in age between monazite and biotite may be as much as 20 m.y.

Of the three total-rock rubidium-strontium ages which could be compared with the uranium-lead data, only one was significantly different from the monazite age. The difference is attributed to a lack of equilibration of strontium during magma genesis and a consequent inheritance of an 'isochron' from the source. The total-rock age is therefore a younger limit to the age of the granitoid's source.
CHAPTER 7

THE FINISTER GRANODIORITE

7.1 THE GRANITOID

The Finister Granodiorite forms a relatively small pluton (30 km$^2$ in area) near the centre of the Berridale Batholith (Fig. F.1). It has been described in detail by White et al. (1978). The Finister is unusual in that it appears to be an integral part of an otherwise metasedimentary screen which is the country rock to several other major plutons. If this interpretation is correct, the implication is that the Finister pluton is older than many, if not all, of the other plutons in the region. In addition, the screen in which the Finister occurs is locally coincident with the I-S line, which White et al. (1976) consider to mark 'a very significant change in the character of the rocks occurring at depth' and to be 'a major tectonic feature of eastern Australia' (p. 111). The Finister is, for these reasons, of considerable interest both geochronologically and geochemically.

The Finister is a hornblende-bearing I-type granitoid which, with the Merumbago Granodiorite (Fig. F.1), has been distinguished by White et al. (1978) as being chemically and petrographically distinct from the other I-type granitoids of the Berridale and Kosciusko Batholiths. The Finister and the Merumbago constitute the 'Finister Suite'. The rock is mafic and medium-grained. Both in hand specimen and in thin section it has a very irregular appearance due to considerable alteration (which is not an effect of weathering) of the primary mineralogy. The hornblende grains are ragged and two generations of hornblende are present. The cores of the grains are brown while the rims, and smaller aggregates, are blue-green. The quartz forms granular aggregates in which triple junctions are common. The calcic cores of the plagioclase grains are sericitised and epidotised. The biotite is chloritised on the grain boundaries and commonly is chloritised and epidotised along the cleavage traces. In many cases finer biotite mantles the larger grains.
It is possible that this alteration is a consequence of the Finister's close proximity to large, later plutons but this is not thought to be so; the Merumbago has an identical texture to that of the Finister, even to the extent of having two generations of hornblende, yet it occurs in relative isolation (Fig. F.1). The alteration is considered, therefore, to be a primary feature of the Finister and the Merumbago granitoids.

White et al. (1978) have cited several features in support of their argument that the granitoids of the Finister Suite are chemically distinct from the other I-type granitoids in the region. These include the fact that, relative to the other I-type granitoids, for a given silica content the granitoids of the Finister Suite are poor in Ti and Na and rich in the ferromagnesian and related elements, Mg, Fe, Se, V, Cr, Co and Ni. Their interpretation of these chemical features is that the granitoids of the Finister Suite have been derived by the partial melting of rocks which are much less fractionated than those which are the source of the other I-type granitoids.

7.2 THE ZIRCON

The zircon sample of the Finister (74-335) was collected from the same locality as geochemical sample BB163 (Fig. F.1). In thin section the zircon is seen to be distributed evenly through all the principal mineral phases, similarly to the distribution of zircon in the Tara. The distribution in the Finister is significantly different from that in the Tara however, in that zircon is present in hornblende (Fig. 7.1). Both in thin section and in grain mounts, the zircon in the Finister is seen very clearly to be strongly corroded (Plate XIII). The grains are equidimensional or prismatic with very low length to width ratios (less than 3:1 in most cases). Some grains have well developed pyramidal or multifaceted crystal terminations, but the majority appear to have had their terminations partially or completely rounded (Plate XIII(a), (b)). In addition some grains have crystal faces developed on one end and are rounded on the other (Plate XIII
Fig. 7.1 The distribution of zircon in the minerals of the Finister Granodiorite, as observed in thin section (for further explanation, see the caption to Fig. 3.1).
FIGURE 7.1
Plate XIII  Light photomicrographs of zircons from the Finister Granodiorite.

Scale subdivisions are 10μ. The fractions illustrated are:

(a)  45μ (5)  (b)  75 - 100μ (5)
(b)  45 - 75μ (3)  (d)  + 100μ (5)
(e)  +100μ (5)  (f)  75 - 100μ (5)
Plate XIV  Light photomicrographs of zircons from the Dalgety Sandstone.

Scale subdivisions are 10μ. The fractions illustrated are:

(a) 60 - 75μ  Purple  (b)  - 60μ
(c) 60 - 75μ  Purple  (d) 60 - 75μ  Purple
(e) 75 -100μ  Pink  (f) 75 -100μ  Pink
Plate XV  Scanning electron photomicrographs of zircons from the Dalgety Sandstone. Scale subdivisions are 30\(\mu\). The zircons are from the mixed 60 – 75 \(\mu\) fraction.
Because no examples of the characteristic pitted surface textures of the sandstone zircon (Plates XIV, XV) were observed, this rounding is considered to be indicative of a lack of chemical equilibrium between the zircon and its environment late in the zircon's history, rather than the result of mechanical abrasion. Fine, needle-like inclusions are very common in the zircons but few examples of zircon cores were found. Several examples of parallel growth or zircon outgrowths were observed (Plate XIII (a)).

7.3 GEOCHRONOLOGY

The Finister is in intrusive contact with two other plutons, the Numbla Vale and the Buckleys Lake, both of which were emplaced relatively early in the history of the Batholith (Fig. 2.1). The Numbla Vale is demonstrably younger than the Finister; accidental xenoliths of Finister Granodiorite which occur within the Numbla Vale already have been discussed above. The relative ages of the Finister and the Buckleys Lake are not as clear but, despite the fact that the boundary between the plutons is convex towards the former (Fig. F.1), White et al. (1978) have concluded that the Finister is the older of the two. The emplacement ages of both the Numbla Vale and the Buckleys Lake are younger limits therefore to the emplacement age of the Finister. The weighted mean of the biotite rubidium-strontium and potassium-argon ages for the Buckleys Lake is 417.2 ± 0.9 m.y. and for the Numbla Vale is 417.9 ± 1.7 m.y. As these ages are younger limits to the emplacement ages of those plutons, they are also younger limits to the emplacement age of the Finister.

The country rock to the Finister has not been differentiated from the Adaminaby Beds. Therefore, although no fossils have been found in the immediate vicinity of the pluton, the country rock is considered to be Eastonian (possibly Bolindian) or older. Its usefulness as a constraint on the Finister's oldest possible emplacement age is limited. If the sedimentary rocks are as old as Early Ordovician for instance, emplacement of the
Finister in the Early Ordovician physically would be possible, but for this to have occurred in what would have been the base of an accumulating sedimentary pile without producing any obvious changes in the pattern of sedimentation would be most unlikely. Nevertheless, similar lithologies are found in the Adaminaby Beds from Bolindian to at least as old as Gisbornian in age. Pending a more detailed study of the Adaminaby Beds, all that can be said definitely is that the Finister may be older than Gisbornian but probably is younger than Bolindian, that is, probably is younger than $460 \pm 10$ m.y. (see section 4.3.4).

### 7.3.1 Rubidium-strontium and potassium-argon analyses

The geochronology of the Finister has been studied by a number of techniques. Compston et al. (1977a) analysed both total-rock samples and mineral separates for rubidium-strontium, Tetley (pers. comm., 1977) analysed biotite and hornblende for potassium-argon and Williams et al. (1976) analysed biotite for rubidium-strontium. Moreover, several additional biotite rubidium-strontium determinations are included in the present study.

Thirteen total-rocks from the Finister were studied, one of which was a mafic xenolith and one of which was hornblende-free. When the rubidium-strontium data were regressed, the scatter about the isochron was found to be in excess of the experimental error ($\text{M.S.W.D.} = 40.7$). No outlier in the population could be detected at the 5% level; exclusion of the xenolith and alternatively of the hornblende-free sample resulted in the M.S.W.D. becoming 38.1 and 23.6 respectively. It was decided nevertheless, that because major and trace element analyses had shown that the hornblende-free sample was significantly different from the other samples of the granitoid (Chappell, pers. comm., 1977), it was best omitted from the calculation of the isochron. This having been done, the scatter remained in excess of the experimental error so it must be modelled before an estimate of the total-rock age can be obtained.

The regression analysis detected both components of scatter which
are dependent upon, and independent of, the samples' rubidium to strontium ratios so Model 4 is considered to be the most appropriate. To select the correct model is in fact not critical for the calculated age is not strongly model dependent (Table 7.1).

Table 7.1

<table>
<thead>
<tr>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
<th>Model 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>(509.7 ± 7.9)</td>
<td>504.6 ± 30.5</td>
<td>508.3 ± 39.8</td>
<td>508.3 ± 39.4</td>
</tr>
</tbody>
</table>

The Finister's total-rock rubidium-strontium age is considered to be 508.3 ± 39.4 (t s.e.) m.y.

The discrepancy between this age and the proposed maximum age of the sedimentary country rock (460 ± 10 m.y.) implies either that the stratigraphic control has been misinterpreted or that the total-rock rubidium-strontium age is not the Finister's emplacement age. Failing an exhaustive search of the country rock in the vicinity of the Finister for fossils, or a detailed study of the stratigraphy of the Adaminaby Beds, little more can be done to improve the stratigraphic control. The possibility that the Finister is not intrusive, but a 'window' in the sedimentary sequence has been considered but rejected following a verification that the country rock adjacent to the pluton is contact metamorphosed. The problem remains one which must be solved isotopically.

Two biotite rubidium-strontium ages measured on samples from the Finister, the mean of which was 417.7 ± 1.9 m.y., were reported by Williams et al. (1976). Subsequent analyses of other biotite samples have not changed this result significantly (the current mean biotite rubidium-strontium age of the pluton is 416.8 ± 3.0 (t s.e.) m.y.) but the spread of the biotite ages now exceeds experimental error (Appendix D). In this respect, the Finister is unique in the Berridale Batholith, and the
possibility that the samples really are of different ages or that the biotite ages have been reset to different extents must be considered.

The biotite samples for which potassium-argon ages have been determined include those samples which have biotite rubidium-strontium ages that have the highest deviations from the mean age, but no excess scatter in the potassium-argon ages has been found. In addition, there is no obvious correlation between the biotite ages measured by the two techniques (Fig. 7.2). This being the case, the biotite ages measured by potassium-argon and those measured by rubidium-strontium cannot be pooled. The mean biotite age is therefore $416.8 \pm 3.0$ (t s.e.) m.y. by rubidium-strontium and $417.4 \pm 2.9$ m.y. by potassium-argon.

Because of the demonstrated susceptibility of biotite ages to resetting (Hanson & Gast, 1967; Hart et al., 1968 and this study), the probability that the Finister's biotites record the emplacement age of the adjacent Numbla Vale and/or Buckleys Lake plutons, rather than that of the Finister, is very high. Attempts have been made to test this possibility by the analysis of more resistant isotopic systems, rubidium-strontium from hornblende and feldspars, and potassium-argon from hornblende.

Compston et al. (1977a) selected sample 74-335 for mineral rubidium-strontium analyses because, not only was it one of the freshest samples available but also it had come from a locality which was further from the adjacent plutons than any other (Fig. F.2). The data which Compston et al. (1977a) obtained on plagioclase, potassium feldspar and hornblende are not fitted to an isochron to within analytical error and an analysis of the excess scatter suggests that Model 3 is the most appropriate. The age calculated using Model 3 is $413.9 \pm 15.3$ (t s.e.) m.y. If the analysis of the total-rock is included in the regression, the excess scatter is not increased significantly but it is modified such that Model 4 becomes the most appropriate. The age is not significantly different from the above ($415.9 \pm 13.6$ (t s.e.) m.y.). The age calculated from the mineral isochron is not detectably different from the biotite rubidium-strontium age but is
Fig. 7.2  The relationship between biotite ages measured by potassium-argon and rubidium-strontium on samples of the Finister Granodiorite. There is no obvious correlation between the ages obtained by the two techniques. The range of biotite rubidium-strontium ages is in excess of analytical error. The estimates of analytical error are two standard errors.
FINISTER GRANODIORITE
BIOTITE AGES

FIGURE 7.2
significantly younger than the total-rock age at the 1% level. If the biotite age has been reset, either the mineral age has been reset also or the time of closure of the minerals and the age of biotite resetting cannot be resolved.

Through the analysis of potassium-argon from hornblende, a more rigorous test for resetting was applied. As has been discussed above, the potassium-argon system in hornblende closes at a temperature at least 200°C higher than it does in biotite. If the resetting of the biotite age of the Finister were a low-temperature process, the hornblende ages may not have been reset or may only partially have been so. The ages which have been measured on hornblende from two localities are not detectably different from one another nor from either of the mean biotite ages at the 5% level. The mean hornblende potassium-argon age of the Finister is 417.2 ± 5.2 m.y. The implication of the mutual agreement of all the mineral ages is that the Finister has cooled rapidly at about 417 m.y., either after it was emplaced or after it was reheated to more than 500°C, probably by the emplacement of the adjacent plutons.

7.3.2 Uranium-lead analyses

In an attempt to clarify the geological significance of the total-rock rubidium-strontium age, zircon from the Finister was analysed for uranium-lead. This aim was not achieved however. The zircon discordance pattern for the Finister is very similar to that from the S-type granitoids and indicates the presence of at least two zircon components with different primary ages (Table 7.2, Fig. 7.3). There is the same difficulty in interpreting the data as was faced when considering the data from the S-type granitoids; it is not clear whether a simple mixing model should be applied or whether a model combining mixing and recent lead loss is more realistic.

A regression analysis of the data showed that they are not fitted to a straight line to within analytical error and that the most appropriate model for the excess scatter is Model 3. If a simple mixing model is to be
Table 7.2  Uranium-lead analytical data from the Finister Granodiorite (74-335). The estimates of precision were calculated as outlined in Appendix C. Common lead was assumed to have the same composition as the lead in potassium feldspar (Appendix E).
### TABLE 7.2
ANALYTICAL DATA FOR THE FINISTER GRANODIORITE (74-335)

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206(^{\text{Pb}}) * (204(^{\text{Pb}}) measured)</th>
<th>ATOM PERCENT RADIOGENIC LEAD</th>
<th>*ATOMIC RATIOS</th>
<th>APPARENT AGES (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>206(^{\text{Pb}})</td>
<td>207(^{\text{Pb}})</td>
<td>208(^{\text{Pb}})</td>
</tr>
<tr>
<td>59/ +100(\mu) (5')</td>
<td>7.32</td>
<td>37.46</td>
<td>504.95</td>
<td>1845±9</td>
<td>84.00</td>
<td>4.84</td>
<td>11.16</td>
</tr>
<tr>
<td>26/ +100(\mu) (5)</td>
<td>25.50</td>
<td>49.60</td>
<td>624.47</td>
<td>715.4±5.1</td>
<td>84.68</td>
<td>4.93</td>
<td>10.39</td>
</tr>
<tr>
<td>28/ 75-100(\mu) (5)</td>
<td>18.40</td>
<td>57.73</td>
<td>775.00</td>
<td>772.3±4.3</td>
<td>84.60</td>
<td>4.85</td>
<td>10.55</td>
</tr>
<tr>
<td>30/ 45-75(\mu) (5)</td>
<td>10.50</td>
<td>72.06</td>
<td>960.37</td>
<td>878.7±1.6</td>
<td>84.48</td>
<td>4.80</td>
<td>10.72</td>
</tr>
<tr>
<td>29/ 45-75(\mu) (3)</td>
<td>9.45</td>
<td>128.26</td>
<td>1520.90</td>
<td>260.4±0.4</td>
<td>85.14</td>
<td>4.74</td>
<td>10.12</td>
</tr>
<tr>
<td>61/ -45(\mu) (3)</td>
<td>0.25</td>
<td>116.91</td>
<td>1546.00</td>
<td>532.6±1.7</td>
<td>84.81</td>
<td>4.75</td>
<td>10.44</td>
</tr>
</tbody>
</table>

*Precision estimates expressed as one standard error.*
Fig. 7.3 A concordia plot of the uranium-lead data from the Finister Granodiorite (Table 7.2). The estimates of analytical error are two standard errors. Line A is the line of best fit to the zircon data and intersects the concordia at 401.2 ±13.0 m.y. and 1087 ±38 m.y. respectively.
FIGURE 7.3

FINISTER GRANODIORITE (74-335)

\[ ^{207}\text{Pb}/^{235}\text{U} \]
applied therefore, more than two end-members must be considered. The intercepts of the line of best fit to the data (Fig. 7.3 Line A) with the concordia are $401.2^{+13.0}_{-79.2}$ m.y. and $1087^{+381}_{-447}$ m.y. respectively. The very large uncertainties are a consequence of the relatively small dispersion in the uranium to lead ratio which was achieved, despite the fact that the uranium concentration in the fractions separated ranged from 505 to 1546 ppm. The total range of 33 m.y. in $^{206}_{238}$Pb/$^{238}$U ages for the zircon fractions from the Finister is considerably less than the range observed in the Maffra for example (144 m.y.), for which the range of uranium concentrations is comparable (780 to 1819 ppm).

7.4 INTERPRETATION OF THE GEOCHRONOLOGY

Because of the inherited component in the Finister zircon, it is very difficult to establish to what extent the zircon has been affected by recent lead loss. The coincidence of the potassium-argon and rubidium-strontium ages (with the exception of the total-rock age) at approximately 417 m.y. is an indication either of a major thermal event at that time or of eventual cooling then, after possibly a series of events. It is necessary to establish therefore, whether or not approximately 417 m.y. ago was a significant time in the history of the zircon.

If the difference between the Finister's biotite age (measured by either potassium-argon or rubidium-strontium) and the age corresponding to the lower intersection of the discordia and the concordia is tested for significance, the two ages are found not to be detectably different from one another at the 5% level. There is thus no evidence in the zircon data for the uranium-lead systems having been disturbed at any time later than that indicated by the other isotopic techniques, namely, there is no evidence for recent lead loss. In that case, discordance predominantly due to the mixing of two components, one $401.2^{+13.0}_{-79.2}$ m.y. old and the other $1087^{+381}_{-447}$ m.y. old cannot be rejected summarily. At the same time, the absence of a detectable difference between the ages does not confirm that such an interpretation is
If the discordance pattern is considered to be a mixing line, at least two means by which it might have been generated must be considered. It may be the result of the severe loss of lead about 400 m.y. ago from zircon which has a primary age of approximately 1087 m.y. Alternatively, the zircon may have crystallised for the most part about 400 m.y. ago, but mixed with it may be inherited zircon 1087 m.y. old. The data do not permit a distinction to be made between these alternatives, and some combination of the two may in fact be closest to the truth. Several observations support this conclusion.

First, the zircons in the Finister are distributed randomly through the principal minerals of the rock (Fig. 7.1). The Finister magma is thought to have contained a relatively high proportion of melt (White et al., 1978), so this distribution is compatible with much of the zircon having been present in the melt and then having been incorporated into minerals at random on quenching, rather than with the crystallisation of zircon as the magma cooled after emplacement.

Secondly, the zircons appear to have been unstable in the magma when it crystallised for the last time. Many of the grains are visibly corroded and few of even the finest grains are euhedral (Plate XIII). This may reflect the zircons' having suffered a complex magmatic history, or a late change in their environment (possibly at the time the Buckleys Lake and Numbla Vale plutons were intruded) which caused them to become unstable. It does argue however, against the age of the lower intercept simply being a time of new zircon growth.

Thirdly, there is a very strong inverse correlation between the $^{207}$Pb/$^{206}$Pb ages of the zircon fractions and their uranium contents (Fig. 7.4). Whatever the process which caused the discordance, it was one which was sensitive to the zircons' chemical composition. That being so, an episode of lead loss from one population must be preferred to the random mixing of two components of unrelated ages and chemistries (unless both
Fig. 7.4 The relationship between the $^{207}\text{Pb}/^{206}\text{Pb}$ age of, and the concentration of uranium in zircon fractions from the Finister Granodiorite. There is a strong inverse correlation between the two parameters. The estimates of analytical error are two standard errors.
FIGURE 7.4

FINISTER GRANODIORITE ZIRCON

\[ 207 \text{Pb}/206 \text{Pb age (m.y.)} \]

Uranium (ppm)
components had a $^{207}\text{Pb}/^{206}\text{Pb}$ - uranium correlation which was strongly developed).

Independently of any theories about the reasons for the discordance, if the zircons are assumed to have suffered no more than a recent disturbance of their lead to uranium ratios since the dispersion in $^{207}\text{Pb}/^{206}\text{Pb}$ was generated, the $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the fractions can be used to place some limits on the times of zircon crystallisation. Some of the zircon must have crystallised at or before $539.2 \pm 13.1$ m.y. (the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ age measured) and some of the zircon must have crystallised (or have lost radiogenic lead) at or after $437.8 \pm 10.9$ m.y. (the youngest $^{207}\text{Pb}/^{206}\text{Pb}$ age measured).

In the light of these results, the total rock age of $508.3 \pm 39.4$ (t s.e.) m.y. is no longer an age in isolation. On the other hand, its geological significance has not been clarified and important aspects of the Finister's history, such as its emplacement age and the age of its source, remain unresolved.

7.4.1 A geochronological model

The source of the Finister magma was similar to, but much less fractionated than, the source of the other I-type granitoids. It formed possibly as long as 1087 m.y. ago, certainly no later than about 539 m.y. ago, or at that time zircon crystallised. About 508 m.y. ago the strontium isotopes in the source were homogenised, probably by a melting event, and the total-rock rubidium-strontium system from that time remained closed, despite a later re-equilibration of strontium between minerals on a local scale. A possible explanation for this is that the melt formed by a relatively high degree of partial melting, crystallised very rapidly soon after separating from the source, and thereafter was mobilised by very small degrees of partial melting and always as one mass so that strontium equilibration on a large scale never again took place.

The time at which the magma was emplaced into its present
position cannot be determined with certainty. It must have been later than 470 m.y. (the maximum possible age for the youngest units in the country rock sedimentary sequence) but earlier than about 417 m.y. (the postulated emplacement ages of the adjacent plutons by which the Finister is intruded). For a reason which is not presently understood, at or before the latter time, the minerals in the magma (or the granitoid) became unstable and were altered and recrystallised.

The Finister was strongly affected by the emplacement of the Buckleys Lake and the Numbla Vale plutons. The potassium-argon systems were completely reset (the temperature probably exceeded 500°C), as was the rubidium-strontium system in most minerals. The only evidence that resetting in fact occurred is seen in the biotite rubidium-strontium ages where small, but significant, differences between the ages of different samples presently are observed.

7.5 SOME POSSIBILITIES FOR FURTHER RESEARCH

There are several avenues of research by which the problem of the age of the Finister may be solved.

1) Zircon can be extracted from a Finister xenolith in an attempt to bypass the effects of inheritance.

2) The second granitoid in the Finister Suite, the Merumbago Granodiorite, can be examined more closely. It is well controlled stratigraphically (there are Eastonian graptolites in its contact aureole) and it should be sufficiently far from later plutons to have escaped resetting. Analysis of zircon from the Merumbago might be particularly valuable in establishing the age of the source of the Finister Suite. Hornblende and biotite potassium-argon ages may enable the rate of the Merumbago’s cooling to be estimated or alternatively, may establish the time at which the alteration of the granitoid took place. The total-rock rubidium-strontium age of aplite which is common in the Merumbago would provide a younger limit to its emplacement age.
3) Single crystal analyses of zircons from the Finister might help in the identification of important events in the zircons' history (the time of primary crystallisation of the older component, of a later episode of lead loss, or of later zircon crystallisation, for instance). Also, episodic lead loss could be distinguished from multicomponent mixing.

4) Ion microprobe lead-lead (or uranium-lead) analyses might be used to establish the growth history of individual grains and to identify zircon (zones or grains) which crystallised in the source or at the time the Finister was emplaced.

7.6 SUMMARY

Geologically, the Finister Granodiorite is one of the oldest plutons in the Berridale Batholith. Its geochronology is complex and remains partially unresolved.

The pluton's total-rock rubidium-strontium age (508.3 ± 39.4 (t s.e.) m.y.) is significantly older than its biotite rubidium-strontium age (416.8 ± 3.0 (t s.e.) m.y.), its age calculated from a mineral rubidium-strontium isochron (415.9 ± 13.6 (t s.e.) m.y.), its biotite potassium-argon age (417.4 ± 2.9 m.y.) and its hornblende potassium-argon age (417.2 ± 5.2 m.y.). The zircon data from the Finister indicate the presence of an inherited component with a minimum primary age of 539.2 ± 13.1 m.y. The scatter in the biotite rubidium-strontium ages measured on different samples exceeds the analytical error.

As the Finister is an I-type pluton and is thought to be derived from an igneous source, 539.2 ± 13.1 m.y. is considered to be a minimum estimate for the age of the source. Because of the likelihood that the pluton's potassium-argon and rubidium-strontium ages (with the exception of the total-rock age) have been reset, the pluton's emplacement age is not considered to have been resolved. An older limit to the pluton's emplacement age, provided by the stratigraphic constraints, is considered to be 470 m.y. A younger limit to its emplacement age, provided by the biotite age of adjacent plutons by which it is intruded is 417.2 ± 0.9 m.y.
CHAPTER 8

THE GEOCHRONOLOGY OF THE BERRIDALE BATHOLITH

The principal finding of the present study is that the ages of any one pluton in the Berridale Batholith measured by the various isotopic techniques commonly are significantly different from one another. Because the differences depend as much upon the minerals analysed as they do upon the decay scheme used, and because it has been demonstrated that agreement between the results of the different techniques is obtained under ideal conditions (Tetley et al., 1977), the differences are considered to be significant geologically, rather than to be a consequence of errors in the decay constants of uranium, rubidium and potassium.

The zircon data from all the granitoids are discordant. The pattern of discordance corresponds closely to the classification of the granitoids as I- or S-type on the basis of the inferred nature of their source materials (Chappell & White, 1974). The S-type granitoids, which are derived by the partial melting of sedimentary rock, contain a zircon component which is considerably older than the stratigraphic limit to the plutons' maximum emplacement age, $460 \pm 10$ m.y. The amount of inheritance in particular granitoids is independent of the amount of mafic solid source residue which the magma contained. In the I-type granitoids, which are derived by the partial melting of igneous rock, the patterns of discordance are more variable.

The two mafic I-type granitoids which were sampled also contain a relatively old inherited component, although their discordance patterns are different from, and the measured $^{207}\text{Pb}/^{206}\text{Pb}$ ages generally are lower than those in the S-type granitoids. The effects of inheritance are not observed in zircon analyses from the two felsic I-type granitoids which were sampled. The fact that the latter two granitoids are the only ones in which the zircon population does not include an abundance of stubby grains with multifaceted terminations possibly is significant.

Because of the absence of evidence for large-scale assimilation
of country rock by any of the granitoids, the zircon is considered to be derived both from the granitoids' source rocks and by crystallisation from the magmas. It is considered that the old zircon components in the mafic I-type granitoids are remnant from an earlier phase of the magmas' histories but that the old component in the S-type granitoids represent detrital zircon from those granitoids' metasedimentary source rocks.

The proposal that zircon may survive magma genesis is supported by the results of many studies of zircon uranium-lead ages done by other workers, notably those in Western Europe, and is not in conflict with the small amount of experimental work which has been done on the stability of zircon in rocks at high temperatures. Zircon has been observed to recrystallise in high-grade metamorphic rocks (Gastil et al., 1967; Davis et al., 1968; Marshall, 1968 and Zabiyaka & Makhlayev, 1970, for example) and at relatively low temperatures in the laboratory (Mumpton & Roy, 1961), but temperatures in excess of 1500°C are required before it begins to dissociate to ZrO₂ + SiO₂ (Mumpton & Roy, 1961; Garcia Verduch & Del Olmo Guillen, 1970; Dyda, 1971). The melting point of zircon from granitoids has been estimated to be 1655 - 1685°C by Dyda (1971), although the zircon dissociation products do not become completely molten until the temperature reaches approximately 2500°C (Garcia Verduch & Del Olmo Guillen, 1970). In view of the abundant corroded zircon which is found in the granitoids of the Berridale Batholith, it is concluded that zircon may break down chemically in granitic magmas at temperatures which are considerably less than its dissociation point.

The effects of lead loss (either by continuous diffusion or during a recent episode) are observed in several of the zircon discordance patterns from the granitoids of the Berridale Batholith. Similar effects are considered to be present in all the other discordance patterns, although they are dominated by the effects of inheritance. As a consequence, the geological significance of the two intercepts of each discordia with the concordia is questionable. The lower discordia intersection age of each of the S-type
granitoids is equal to, or significantly younger than, the ages measured on the same granitoid by biotite potassium-argon, biotite rubidium-strontium or monazite uranium-lead. The lower intersection age is not considered to be significant geologically; although it is in fact a younger limit to a pluton's emplacement age, emplacement can be dated more accurately by other techniques. The discordia upper intersection ages possibly indicate the provenance age of zircon in the granitoids' source rocks (the range of upper intersection ages for the S-type granitoids is relatively narrow, 1343±213 m.y. to 1926±76 m.y.) but there are insufficient zircon data available from the Snowy Mountains region at present for it to be possible to argue for the geological significance of these ages on the grounds of their regional constancy. The oldest $^{207}\text{Pb}/^{206}\text{Pb}$ age measured on zircon from an S-type granitoid is 1163 ± 5 m.y., which is a younger limit to the provenance age of the oldest zircon in the granitoids' source rocks. Those source rocks are considered to be metasedimentary basement which underlies the Snowy Mountains region but which has no local surface equivalent.

Thirty biotite samples from the granitoids of the Berridale Batholith were dated by rubidium-strontium and without exception, the biotite rubidium-strontium and potassium-argon ages determined on each sample are not detectably different from each other. For only one pluton are the differences between the rubidium-strontium biotite ages of different samples significantly greater than the experimental error. Significant differences are common however, between the biotite ages of the plutons. Because these differences in all cases are in agreement with the intrusive sequence determined geologically, the oldest biotite age measured on a pluton is considered to be a younger limit to that pluton's emplacement age. In the Berridale Batholith, the biotite ages are equal to or significantly older than the lower intersection ages of the zircon discordias and are therefore more accurate estimates of the plutons' emplacement ages than the latter.

With a single exception, in which the biotite and monazite ages of a granitoid agree, the monazite age of each granitoid is significantly
older than its biotite age. The monazite ages are considered to be monazite crystallisation ages and are therefore equal to or older than the plutons' emplacement ages. This being so, the monazite age of each granitoid is a younger limit to the age of the latest melting event in that granitoid's history. The differences between the monazite and biotite ages of plutons from the Berridale Batholith range from 0 to $21.9 \pm 3.5$ m.y.

Only a small amount of information presently is available on the blocking temperature of monazite or on its stability at high temperatures. Köppel and Grünenfelder (1975) concluded from a study of monazite ages in a region of high-grade metamorphism that monazite may retain a significant proportion of its radiogenic lead at temperatures as high as 600°C. Ghouse (1968) reported experiments on structural changes in monazite when it was heated to 1130°C for prolonged periods; this temperature is therefore a lower limit to monazite's melting point. It is concluded that there is no evidence of the inability of monazite to survive magmatic temperatures which would preclude the postulate that monazite crystallises and that its uranium-lead system closes prior to granitoid emplacement.

The granitoids' total-rock rubidium-strontium ages are equal to or older than their biotite ages. In those cases in which they are older, it is considered that the total-rock age either is an indication of the evolution of the granitic magma over an extended period of time, or is due to the inheritance of an isochron from the granitoid's metasedimentary source rocks (Roddick, 1974). In view of experiments which have shown that strontium in felsic rocks diffuses very slowly and equilibrates by diffusion only over a range of tens of centimetres at magmatic temperatures (Misra & Venkatasubramanian, 1977 and Magaritz & Hofmann, 1977, for example) it would be possible for heterogeneity in a magma's strontium isotopic composition to be maintained (and pre-magmatic total-rock rubidium-strontium ages consequently to be preserved) as long as the magma was not physically mixed. It is concluded that although total-rock rubidium-strontium ages are not in
all cases the age of an event, they are a younger limit to the age of a granitoid magma's source.

The biotite ages of plutons from the Berridale Batholith, the best estimates of the plutons' minimum emplacement ages, have been related to the stratigraphic time scale by Williams et al. (1976). The ages indicate that the time of cooling of the plutons to the biotite blocking temperature ranges from Late Silurian to Early Devonian. One of the monazite ages is within this range, but the remainder are Mid to Early Silurian or Late Ordovician, if the estimate of the age of the Ordovician-Silurian boundary of Lanphere et al. (1977) is used (435-437 m.y.). Some of the zircon ages which have been interpreted as magmatic zircon crystallisation ages also are Ordovician. It is concluded that the magmatic activity which led to the formation of the Berridale Batholith granitoids extended from at least as early as the Late Ordovician to the Devonian but that the granitoids were emplaced into their present positions over a relatively short period of time from the Late Silurian to the Early Devonian.
APPENDIX A

ANALYTICAL TECHNIQUES

A.1 URANIUM-LEAD FROM ZIRCON AND MONAZITE

A.1.1 Crushing

The samples were broken into blocks less than 2 kg in weight using a sledge hammer and mild steel anvil. All external surfaces, including those which were freshly broken in the field, were discarded. The blocks were examined for veining, weathering and jointing and rejected if any of these was observed. They were then blown with compressed air to remove surface dust.

The blocks were reduced to chips less than 5 mm in grainsize in a jaw crusher then to a fine powder in a Christy & Norris Ltd. overhung disintegrator. The disintegrator was a 28 cm (11") hammer mill in which the sample's grainsize was reduced by impact until it passed a 2 mm screen that lined one side of the crushing chamber. This method of crushing had the advantage that the rock was disaggregated but not ground. Approximately 95% of the resulting powder was finer than 1 mm.

A.1.2 Mineral separation

The sample was slimed and a preliminary heavy mineral concentrate prepared using a Wilfley shaking table. Those samples for which the zircon yield was expected to be low were passed over the table twice. The table was adjusted for maximum zircon yield rather than for maximum efficiency in the concentration of heavy minerals, so samples processed in this way were reduced in weight only by a factor of about ten.

After the concentrate had been dried, it was processed through a heavy-liquid infracentrifuge (Fig. A.1). The infracentrifuge was constructed at Earth Sciences using elements of the design of Verschure and IJlst (1966) and from photographs very kindly supplied by
Fig A.1 Exploded sketch showing the principal components of the continuous-feed, overflow infracentrifuge constructed from the design of Verschure and IJlst (1966). Sample was fed from a hopper (b), through a funnel (d) into a mixing vessel (e) where it was mixed with s-tetrabromoethane from a storage flask (a) by a stainless steel-Teflon R stirrer (c) to form a slurry. The slurry was fed through a tube (f) to the base of the centrifuge vessel (k), which was mounted on a belt-driven pulley (l) that was rotated at about 400rpm. The slurry which overflowed contained those minerals that were less dense than 2.96 g/cc. The slurry was deflected into the catching vessel (g) by an acrylic collar (i). Spraying of the slurry was prevented by a cover plate (h). Contact between the rotating centrifuge vessel and the stationary catching vessel was prevented by a Teflon R bearing (j). The slurry was fed to a filtration vessel (n). The solid residue was retained on filter paper (o) which was fitted to the base of an inner glass cylinder (m) and supported by stainless steel mesh (p). The s-tetrabromoethane was recycled to the storage flask (a) via a Teflon R tube (r) by a Teflon R-lined, peristaltic pump (q).
OVERFLOW INFRACENTRIFUGE

FIGURE A.1
the latter. The centrifuge was operated with s-tetrabromoethane (density 2.96 g/cc), which was cheaper and less toxic, although also less dense, than diiodomethane (density 3.3 g/cc). The sample was mixed with heavy liquid and fed to the bottom of the centrifuge vessel (which was rotated at approximately 400rpm) so that the latter was overflowed. Minerals more dense than 2.96 g/cc were retained in the vessel and those that were less dense were carried out with the overflow. The overflow was filtered and the heavy liquid recycled. The heavy mineral concentrate obtained from the centrifuge contained less than 1% minerals less dense than 2.96 g/cc.

Because of the differences between individual heavy-mineral concentrates, the mineral separation procedure which followed could not be standardised. In general, the sample was reduced in bulk as much as possible by a series of magnetic separations. A hand magnet was used for strongly magnetic minerals such as magnetite and a Cook isodynamic separator for minerals such as hornblende and biotite. This was not as efficient as a density separation using diiodomethane would have been, but was necessary because of the large amounts of sample which otherwise would have to have been handled. The magnetic separation reduced the sample sufficiently for a diiodomethane separation to be feasible.

Following the diiodomethane separation, the heavy mineral concentrate, in some cases, was dominated by non-magnetic sulphides. If these hampered further separations they were dissolved with warm 1N HNO₃ but only after a concentrate of monazite had been set aside. The zircon concentrate was run through the magnetic separator at increasing magnet currents until zircon began to appear in the residues. The sample was then sized using new nylon mesh held in plastic mounts and a set of magnetic fractions was prepared.

The fractions which were selected for analysis were subjected to a final purification. Washing the sample down a fold in a piece of filter
paper with alcohol was an efficient method of removing residual opaque minerals but in all but a few cases, the final sample had to be prepared by hand picking (the exceptions were those fractions finer than 45μ, for which hand picking was not attempted). The hand picked samples were better than 99% pure.

A.1.3 Chemical processing
A.1.3.1 Dissolution

The method used for the dissolution of zircon and monazites and the extraction of uranium and lead was slightly modified from the low-contamination method of Krogh (1973). All the reagents used were sub-boiling distilled in quartz and/or double-bottle Teflon\textsuperscript{R} (Mattinson, 1971) stills and all the operations except the initial dissolution (during which the sample was isolated in a sealed capsule) and weighing, were carried out in filtered air work stations certified Class 100.

The weight of sample processed ranged between less than 1 and more than 30 mg, depending upon the availability of material and the extent to which the chemical procedures had been refined at the time. The advantages in using small samples were that they could be hand picked quicker and more efficiently than large ones and that selected grains which represented a tiny proportion of the total population could be analysed.

The zircon (but not the monazite) was washed prior to dissolution. The washing was done in the capsule which was to be used for the initial dissolution and after the sample had been weighed. The washing procedure was: five minutes in hot (near to boiling) 5N HNO\textsubscript{3}, two rinses with cold water to remove the acid and dissolved salts, five minutes in hot water. The reagents were changed between washes with a quartz-tipped syringe. The purpose of the wash was to remove surface contamination and residual sulphide which may have adhered to the grains. From the results of experiments by Silver and Deutsch (1963) in which the
effects of a much stronger acid wash (one hour in concentrated HNO₃ at 80°C) were examined, it was concluded that the adopted washing procedure would have no deleterious effect on the zircons themselves.

Both the zircon and monazite were dissolved in a high-pressure Teflon R bomb of Krogh's (1971) design. The dissolution reagent for monazite was 1.5 ml of 6.2N HCl mixed with about 0.2 ml of 7N HNO₃. Monazite dissolved overnight when the bomb was heated to 200°C. The dissolution reagent for zircon was 1.5 ml of HF mixed with 0.2 ml of 7N HNO₃. Zircon dissolved when the bomb was maintained at 200°C for seven days.

A.1.3.2 Spiking

When the dissolution was complete, the dissolution reagent was dried off and replaced with 2 ml of 3.1N HCl. The Teflon R bomb liner was reassembled without its outer sleeve and maintained at approximately 80°C on a hotplate overnight. The solution was then cooled and the moisture which had condensed on the lid thoroughly remixed with it.

The solution was split into two parts in an accurately measured proportion of about 2:1. Mixed ²⁰⁸Pb-²³⁵U tracer was added to the smaller part, which was then heated to 200°C in the bomb overnight. The unspiked aliquot was retained for the measurement of the lead composition.

A.1.3.3 Extraction of uranium and lead

Uranium and lead were extracted using anion exchange resin. The resin (AG 1-X8 200-400 mesh), which had been prewashed with 1N NaOH and 6N HCl and stored in H₂O, was loaded into a column made from shrink Teflon R tubing and with a polypropylene frit. The internal diameter of the column was about 5 mm so the 0.5 cc of resin used formed a bed about 2.5 cm high. New resin was used for every sample and the columns were washed between samples by being stored in 1N HNO₃. The column procedure was as follows:
1) Load the resin.

2) Wash the resin with 2 mls of 6.2N HCl, 2 mls of H$_2$O, 2 mls of 6.2N HCl and 2 mls of H$_2$O.

3) Condition the resin with 1 ml of 3.1N HCl.

4) Load the sample in 1 ml of 3.1N HCl.

5) Elute with 2 mls of 3.1N HCl (elution with 1 ml of 3.1N HCl as suggested by Krogh (1973) was found to be inadequate).

6) Collect lead in 1 ml of 6.2N HCl.

7) Collect uranium in 1 ml of H$_2$O.

To minimise cross contamination, the unspiked sample was run first, then the spiked sample was run through the same column after the resin had been changed.

The extent of laboratory contamination was assessed from total-processing blanks, one of which was run concurrently with each batch of five samples. The total-lead blank fluctuated apparently at random about a mean value that rose from 1ng to 2ng and fell again over an eighteen-month period. Individual determinations of the blank ranged from 0.2ng to as much as 3.1ng during that time. The blank was attributed primarily to airborne contamination and was corrected for by analysing the composition of dust gathered by the laboratory's air filters (Nieuwland, *pers. comm.*). This was

\[
\frac{^{208}_{\text{Pb}}}{^{206}_{\text{Pb}}} = 2.1452 \pm .0004, \quad \frac{^{207}_{\text{Pb}}}{^{206}_{\text{Pb}}} = 0.9046 \pm .0001, \quad \frac{^{206}_{\text{Pb}}}{^{204}_{\text{Pb}}} = 17.028 \pm .003
\]

The uncertainties are expressed as two standard errors of the mean. 0.2ng was a maximum estimate for the contribution to the blank from reagents. Uranium contamination was much less significant than that of lead and averaged 0.15ng per sample.
A.2  RUBIDIUM-STRONTIUM

A.2.1  Crushing

Total-rock samples for rubidium-strontium were commonly less than 5 kg in weight. The sample was broken into blocks with a sledge hammer, reduced to chips less than 5 mm grain size in a jaw crusher and passed through a roll mill. The resulting powder was divided into portions, each of approximately 50 g, and ground for a short time in a Sieb swing mill, then recombined and mixed thoroughly. A 50 g aliquot was separated by quartering and ground in the swing mill for a further five minutes. Samples from which minerals were to be separated were processed similarly except that the coarse powder from the roll mill was used.

A.2.2  Mineral separation

Biotite was the only mineral separated for rubidium-strontium analysis for the present study. The powder from the roll mill was sized to 350-700μ and a biotite concentrate obtained using a Carpco magnetic separator. The concentrate was washed in an ultrasonic bath to remove dust and the biotite concentrated further using heavy liquid (s-tetra-bromoethane). To break up aggregates and to remove those minerals which were not sheet silicates, the sample was placed on a flat plate and rolled with a metal roller. The dust was removed by sieving and a further wash in an ultrasonic bath. The chlorite that most samples contained was removed using heavy liquids and a Cook isodynamic magnetic separator.

A.2.3  Chemical processing

Biotite samples weighing 50-80 mg were processed. They were dissolved in a mixture of 5 ml of HF and 0.1 ml of HClO₄, then converted to chlorides with a few drops of 6N HCl. The salts were dissolved in 2 to 5 ml of hot 2.5N HCl then the solution made up to about 30 ml with water. To this the mixed ⁸⁵Rb-⁸⁴Sr tracer was added. When the sample-
spike mixture had been dried it was taken up in 1 ml of 0.25M $\text{C}_2\text{H}_2\text{O}_4$ in 0.5N HCl in preparation for the extraction of rubidium and strontium.

Rubidium and strontium were extracted using 2 g (wet weight) columns of cation exchange resin and the oxalic acid technique devised by Dr J.J. Foster (pers. comm.). The resin was conditioned with 5 ml of 0.25M $\text{C}_2\text{H}_2\text{O}_4$ in 0.5N HCl before the sample was loaded. After the sample was loaded, the column was flushed with the $\text{C}_2\text{H}_2\text{O}_4$-HCl mixture until the eluant was no longer coloured by Fe. This took 5-10 ml. The balance of the equivalent of 15 ml 1N HCl was rinsed through the column in four separate washes to remove the excess oxalate. Rubidium was collected in a further 6 ml of 1N HCl, the column was washed with 6 ml of 2.5N HCl and strontium was collected in a further 4 ml of 2.5N HCl.

A second pass through the columns was necessary to purify the samples sufficiently for mass spectrometry. The conventional HCl technique was used for this: the sample was loaded in 1 ml of 1N HCl, the column washed with 15 ml of 1N HCl, rubidium collected in 6 ml of 1N HCl, the column washed with 6 ml of 2.5N HCl and strontium collected in 4 ml of 2.5N HCl.

A.3 SAMPLE LOADING FOR MASS SPECTROMETRY

All the isotopic analyses were done on the MSZ 23 cm, 60° sector field mass spectrometer that was designed by Clement and Compston (1972) and housed at the Research School of Earth Sciences, A.N.U. The samples were loaded as solids on rhenium filaments. Rubidium and strontium were loaded in water as chlorides on the side filaments of a triple filament assembly as described by Compston et al. (1965). Uranium was loaded in 0.2N HNO$_3$ (Tilton et al., 1955) but otherwise, in the same way. The loading of lead was more complex.

Lead was loaded onto a single filament using a modified version of the technique described by Akishin et al. (1957). For the loading, a syringe to which a disposable, clean quartz tip was fitted was used. The
tip of the syringe was wet with 0.2N HNO₃, then used to transfer one drop (about 20μl) of 0.76N H₃PO₄ onto the sample. A small amount of H₃PO₄ (about 2μl) was retained and mixed with a drop of silica gel (about 20μl) and placed on the filament. The filament was heated to just below 100°C. When the water had evaporated from the drop of gel, a second drop was added and allowed to 'dry'. The presence of the small amount of H₃PO₄ prevented the mixture on the filament ever going to complete dryness.

During the thirty minutes which it took to load the gel, the H₃PO₄-sample mixture was agitated periodically with the syringe to ensure that the sample dissolved. The mixture was transferred to the damp gel and the filament temperature increased to slightly above 100°C. This ensured that when the filament temperature was further increased, the solution did not boil.

When the mixture once more had 'dried', it had the appearance of a solution, but as the filament temperature slowly was raised, a point was reached at which a grey or brown precipitate appeared. The temperature was raised further until the H₃PO₄ began to fume then held steady until fuming ceased. The last of the H₃PO₄ was removed by raising the filament to red heat for 1-2 seconds and the loading was then complete.

The H₃PO₄ used for loading was prepared from Ultrex® sublimed P₂O₅. The silica gel was made by mixing solid chromatography-grade Merck silica gel (2.5 g) with water (200 ml) in a silica beaker and agitating the mixture for 5 minutes with a Sonifier Cell Disruptor B-12 (made by the Branson Sonic Power Co., Connecticut). The mixture was allowed to stand for four hours then decanted. The supernatant was allowed to stand for a further four hours and the decantation repeated. This ensured that the coarse, residual solid gel was removed.
APPENDIX B
DATA REDUCTION

B.1 THE 'ON-LINE' PROCESSING OF MASS SPECTROMETER DATA

The mass spectrometers were linked on-line to a Hewlett packard 2116B computer. It had been programmed by Dr. P.A. Arriens to operate as a multi-user, time-sharing system (Arriens, 1973). The acquisition and reduction of data was performed by a succession of programs which were swapped automatically from disc to core as they were needed. When either rubidium or strontium was measured, corrected isotopic ratios and elemental concentrations were calculated at the conclusion of each run. No similar facility was available for reducing uranium or lead data so the necessary programs were written and installed.

Because the rapid peak switching technique used for the measurement of isotopic ratios of rubidium and strontium was equally applicable to uranium and lead, the data acquisition programs of Arriens were used without significant modification. Consequently, the initial treatment of the raw data as performed by these programs was the same as for rubidium and strontium. The basic philosophy behind this treatment has been outlined by Arriens and Compston (1968) and was as follows.

Beam intensity measurements were made in pairs by rapid switching of the magnetic field (or accelerating voltage) to preselected values, bringing each beam into the collector alternately. A typical data set might consist of twelve to thirty one-second integrations of firstly one beam, then the other, separated by delays of one second to allow switching. Up to sixteen such sets, in which different isotopic ratios were measured, would constitute a run.

Before a best estimate of the ratio determined in each set could be calculated, zero-point corrections were necessary. They consisted of the 'static zero', which was a function of the electrometer range and the 'dynamic zero', a small but significant component which was a function of
The static zero, for the most part, was attributable to a constant zero offset potential applied to the electrometer by a mercury cell. This ensured that the voltage to frequency converter was never required to measure less than 10% of full scale deflection and avoided systematic errors which would otherwise have resulted from its non-linearity for zero input. The static zero was determined by taking repeated measurements at non-integral mass stations remote from any tailing effects from the sample. Mass 202.5 was used for lead and mass 238.8 for uranium. A range of settings was monitored during each run and only very rarely were these particular settings found not to give reliable estimates of the background.

The dynamic zero was added to the static zero to correct for the finite response time of the electrometer input circuit. The correction was a machine constant and predetermined by several runs specifically designed for the purpose (see Appendix C). The dynamic correction applied to any one beam was proportional to the size of the other beam which was being used in the calculation of the isotopic ratio.

Within a set, a ratio was calculated for each time a switch was made from one beam, to the other and back again. The set of ratios so obtained was subjected to two rejection tests which were cycled until both were passed. The mean, standard error and coefficient of variation of the set having been calculated, the determination with the highest deviation was identified and tested against a limit usually set at 2.2 standard errors of the population. The coefficient of variation of the population was then tested against an expected maximum value of 1%. In some runs where very large ratios were being measured and signals were small the expected coefficient of variation was considerably larger than 1% so the limit was raised, sometimes to as much as 90%. Failure of either test resulted in the identified determination being rejected and the cycle was begun anew with a freshly calculated mean, standard error and coefficient of
variation. These tests were very efficient in removing the effects of spurious beam fluctuations yet allowed a realistic estimation of the precision of the determination to be made.

At the conclusion of a run, the data available for further calculation consisted, for each set, of the set number, corrected mean ratio, standard error of the mean and the time of the measurement. It was this data which was utilised by subsequent programs.

B.1.1 Uranium data

The reduction of uranium data involved the calculation of a best estimate for $^{235}\text{U}/^{238}\text{U}$ and then calculation of the uranium concentration using spike and spiking data entered by the operator.

A weighted mean $^{235}\text{U}/^{238}\text{U}$ was computed, each determination being weighted inversely as the square of the standard error of the mean. For determinations $m_1$, $m_2$, $m_3$ ... with the respective standard errors of the mean $S_1$, $S_2$, $S_3$ ..., the weighted mean $M$ was given by

$$
M = \frac{\frac{m_1}{S_1^2} + \frac{m_2}{S_2^2} + \frac{m_3}{S_3^2} + \ldots}{\frac{1}{S_1^2} + \frac{1}{S_2^2} + \frac{1}{S_3^2} + \ldots}
$$

The uranium concentration was calculated from the isotope dilution equation in which the moles of $^{235}\text{U}$ in the sample ($^{235}\text{U}_{\text{sa}}$) were related to the moles of $^{235}\text{U}$ in the spike ($^{235}\text{U}_{\text{sp}}$) through the spike $^{238}\text{U}/^{235}\text{U}_{\text{sp}}$ ($^{238}/^{235}_{\text{sp}}$), the sample $^{238}\text{U}/^{235}\text{U}$ ($^{238}/^{235}_{\text{sa}}$) and $^{238}\text{U}/^{235}\text{U}$ in the resulting mixture ($^{238}/^{235}_{\text{m}}$).

$$
^{235}_{\text{sa}} = ^{235}_{\text{sp}} \cdot \frac{(^{238}/^{235})_{\text{m}} - (^{238}/^{235})_{\text{sp}}}{(^{238}/^{235})_{\text{sa}} - (^{238}/^{235})_{\text{m}}}
$$

B.1.2 Lead data

In the reduction of lead data, it was assumed that differences
between repeated determinations of any one isotopic ratio were due to the randomness of ionisation, zero-point noise and systematic, mass dependent, variable mass discrimination (fractionation). The data were collected in a series of 'sub-runs', each consisting of a measurement of $^{208}\text{Pb}/^{206}\text{Pb}$, followed by a number of determinations relating $^{207}\text{Pb}$, $^{206}\text{Pb}$ and $^{204}\text{Pb}$ and terminated by a further measurement of $^{208}\text{Pb}/^{206}\text{Pb}$. During a sub-run, the operating conditions of the mass spectrometer were not altered so the rate of fractionation was assumed to be constant. It was the function of the computer to test both within and between sub-runs for significant fractionation and to correct for it if necessary. At the conclusion of the run, a best estimate for each ratio and its uncertainty based on internal precision was calculated.

In order that the operator might be allowed as much flexibility as possible, a system by which ratios could be identified at the end of a run, rather than a rigid input format, was used. It was possible therefore to relate any isotope to any other as many times as required so long as the notion of sub-runs was adhered to. The system required the operator to identify which isotopes had been measured in each set and which was the numerator and which the denominator in the ratio subsequently calculated.

The first step was to ensure a uniform format. A test was performed on each ratio label to check whether the abundance of the isotope of larger mass was in the numerator. If such was not the case, the mean ratio was replaced by its inverse and its standard error was appropriately adjusted, the coefficient of variation being constant. For example, if a ratio was labeled '206/204' it was left untouched but if the label was '204/206' then 206/204 and the standard error of the mean 206/204 were calculated and substituted for the original values.

The calculation of the revised standard error of the mean relied on the fact that the coefficient of variation of a set was independent of the ratio calculated. In other words, if a set consisted of $n$ determinations of a ratio $x/y$, if the standard error of the mean $x/y$ was $S_{x/y}$ and if the...
Coefficient of variation of the set was $C_v$ then

$$C_v = \frac{S_{x/y}}{\sqrt{\frac{\sum (x/y)}{n}}} = \frac{S_{y/x}}{\sqrt{\frac{\sum (y/x)}{n}}}$$

so

$$S_{y/x} = S_{x/y} \cdot \left(\frac{\overline{y/x}}{\overline{x/y}}\right)$$

In order to do this calculation with the data available, it was necessary to make the approximation that

$$\frac{\overline{y/x}}{\overline{x/y}} = \frac{1}{x/y}$$

The errors introduced by doing so were minimal. The relationship employed was

$$S_{y/x} = S_{x/y} \cdot (1 / \overline{x/y})^2$$

The next operation was the identification of individual sub-runs by searching for the determinations of $^{208}\text{Pb}/^{206}\text{Pb}$ which bounded them. This having been done, each sub-run was tested to determine whether or not there was a significant difference between those bounding determinations, given the calculated internal precision.

The difference between the means ($\delta$) was considered significant at the 5% level when $\delta$ equalled or exceeded $\mu_5 S$. $\mu_5$ was a constant dependent on the standard errors of the two means and the number of observations represented by the means. By assuming the standard errors of the means to be of the same magnitude and taking the numbers of observations to be twelve per set, $\mu_5$ was fixed at 2.13. $\mu_5$ was relatively insensitive to the number of observations; had thirty per set been assumed, $\mu_5$ would have been been 2.0. Values for $\mu_5$ have been tabulated by Lyon, 1970 (p 344).

$S$ was the standard error of the difference. It was a function of the standard errors of the means ($S_1$ and $S_2$) in that

$$S = \sqrt{S_1^2 + S_2^2}$$
The result of the significance test thus depended only upon the two means \((m_1\) and \(m_2\)) and their standard errors. The difference was considered significant if the following was true:

\[
\frac{|m_1 - m_2|}{\sqrt{S_1^2 + S_2^2}} \geq 2.13
\]

Further treatment of the data in the sub-run depended upon the outcome of this significance test. If the fractionation was determined to be insignificant, the best estimate for \(^{208}\text{Pb}^{206}\text{Pb}\) was calculated as a weighted mean of the two bounding determinations of \(^{208}\text{Pb}^{206}\text{Pb}\). For the purpose of calculations to follow, the time ascribed to the mean \(^{208}\text{Pb}^{206}\text{Pb}\) determination was weighted similarly. This same time was ascribed to all other sets in the sub-run.

If the fractionation was considered to be significant, all the ratios in the sub-run were corrected to the time at which the first \(^{208}\text{Pb}^{206}\text{Pb}\) was measured. Fractionation was assumed to have been mass dependent. The first determination of \(^{208}\text{Pb}^{206}\text{Pb}\) was chosen in preference to the second on the premise that it would always be the less susceptible of the two to any fractionation which may have been a function of changes in beam focusing caused by drift of the quadrupole lens potentials or a shift in the point on the filament from which emission was taking place. The correction for mass dependent, variable mass discrimination was carried out as follows.

For a general ratio \(x/y\) (in which \(x>y\)) measured at time \(t_{x/y}\) and bracketed by two determinations of \(^{208}\text{Pb}^{206}\text{Pb}\) made at times \(t_1\) and \(t_2\) respectively (namely, \(t_2 > t_{x/y} > t_1\)), the rate of change of \(^{208}\text{Pb}^{206}\text{Pb}\) as a function of \((^{208}\text{Pb}^{206}\text{Pb})\) \(_t^2\) \((r)\) was given by

\[
r = \frac{(^{208}\text{Pb}^{206}\text{Pb})_{t_2} - (^{208}\text{Pb}^{206}\text{Pb})_{t_1}}{(^{208}\text{Pb}^{206}\text{Pb})_{t_2}} \cdot \frac{1}{t_2 - t_1}
\]
As there was a difference of two mass units between 208 and 206, the fractionation rate per unit mass difference (R) was

\[ R = \frac{r}{2} \]

Given \((\frac{208\text{Pb}}{206\text{Pb}})_{t_2}\) and R, it would thus have been possible to determine \((\frac{208\text{Pb}}{206\text{Pb}})_{t_1}\) by using

\[
(\frac{208\text{Pb}}{206\text{Pb}})_{t_1} = (\frac{208\text{Pb}}{206\text{Pb}})_{t_2} [1-R(208-206)(t_2-t_1)]
\]

Similarly, having calculated R from the rate of change of \(\frac{208\text{Pb}}{206\text{Pb}}\), any ratio \(x/y\) could be normalised to \(t_1\).

\[
(x/y)_{t_1} = (x/y)_{t_{x/y}} [1-R(x-y)(t_{x/y}-t_1)]
\]

(1)

Calculation of the standard error of ratios normalised in this manner involved making an allowance for the fact that the value for R was uncertain, since it was calculated from two determinations of \(\frac{208\text{Pb}}{206\text{Pb}}\) which were themselves imprecise. Expansion of (1) shows all the variables involved in calculating \((x/y)_{t_1}\).

\[
(x/y)_{t_1} = (x/y)_{t_{x/y}} [1- \frac{(208\text{Pb}/206\text{Pb})_{t_2}-(208\text{Pb}/206\text{Pb})_{t_1}.(x-y).t_{x/y}-t_1]}{(208\text{Pb}/206\text{Pb})_{t_2}}]
\]

In the assessment of sources of uncertainty, those variables which could be considered to be without error were grouped with constants to define a constant (K).

\[
(x/y)_{t_1} = (x/y)_{t_{x/y}} [1- \frac{(208\text{Pb}/206\text{Pb})_{t_2}-(208\text{Pb}/206\text{Pb})_{t_1}.K]}{(208\text{Pb}/206\text{Pb})_{t_2}}]
\]

(2)

where \(K = \frac{x-y}{2} \cdot \frac{t_{x/y}-t_1}{t_2-t_1}\).
The function was then left with three variables, \((x/y)\), \((208^{\text{Pb}}/206^{\text{Pb}})\), and \((208^{\text{Pb}}/206^{\text{Pb}})\).

In the general case of a function of three variables \((u = f(x, y, z))\) the standard error of \(u\) \((S_u)\) could be expressed in terms of the standard errors of the variables \((S_x, S_y\) and \(S_z)\).

\[
S_u = \sqrt{\left(\frac{\delta u}{\delta x} \cdot S_x\right)^2 + \left(\frac{\delta u}{\delta y} \cdot S_y\right)^2 + \left(\frac{\delta u}{\delta z} \cdot S_z\right)^2}
\]

Translating this to the problem in hand, for simplicity let \(r_1 = (x/y)\), \(r_{x/y} = (x/y)\), \(R_1 = (208^{\text{Pb}}/206^{\text{Pb}})\), and \(R_2 = (208^{\text{Pb}}/206^{\text{Pb}})\).

\[
S_{r_1} = \sqrt{\left(\frac{\delta r_1}{\delta r_{x/y}} \cdot S_{r_{x/y}}\right)^2 + \left(\frac{\delta r_1}{\delta R_1} \cdot S_{R_1}\right)^2 + \left(\frac{\delta r_1}{\delta R_2} \cdot S_{R_2}\right)^2}
\]

Returning then to (2) and solving the partial derivatives:

\[
\frac{\delta r_1}{\delta r_{x/y}} = 1 - \frac{R_2 - R_1}{R_2} \cdot K
\]

\[
\frac{\delta r_1}{\delta R_1} = -\frac{r_{x/y} \cdot K}{R_2}
\]

\[
\frac{\delta r_1}{\delta R_2} = \frac{r_{x/y} \cdot K \cdot R_1}{R_2^2}
\]

The expression for the standard error of \(r_1\) then became:

\[
S_{r_1} = \sqrt{\left((1 - \frac{R_2 - R_1}{R_2}) \cdot S_{r_{x/y}}\right)^2 + \left(-\frac{r_{x/y} \cdot K}{R_2} \cdot S_{R_1}\right)^2 + \left(\frac{r_{x/y} \cdot K}{R_2^2} \cdot S_{R_2}\right)^2}
\]

By making some assumptions this expression was simplified. The fractional change in \(^{208}\text{Pb}^{/206}\text{Pb}\) through a sub-run was relatively small and consequently, so was the difference between \(R_1\) and \(R_2\).
Hence \[ 1 - \frac{K(R_2 - R_1)}{R_1} \geq 1 \] for \( K \) was usually about 1

and \( \frac{R_1}{R_2} \geq 1 \)

\( S_{r_1} \) was therefore given by

\[
S_{r_1} = \sqrt{\left( (s_{x/y})^2 + \frac{-r_{x/y} \cdot K}{R_2} \cdot S_{R_1} \right)^2 + \left( \frac{r_{x/y} \cdot K}{R_2} \cdot S_{R_2} \right)^2}
\]

namely

\[
S_{r_1}^2 = S_{x/y}^2 + (r_{x/y} \cdot K)^2 \left( \frac{S_{R_1}^2}{R_2} + \frac{S_{R_2}^2}{R_2} \right)
\]

It was possible to simplify the expression still further by assuming that the coefficient of variation of \( R_1 (C_{R_1}) \) could be approximated by \( \frac{S_{R_1}}{R_2} \).

Rearranging (3) gave

\[
S_{r_1}^2 = r_{x/y}^2 \left[ \frac{S_{x/y}^2}{r_{x/y}^2} + K^2 \left( \frac{S_{R_1}^2}{R_2} + \frac{S_{R_2}^2}{R_2} \right) \right]
\]

which simplified to

\[
S_{r_1}^2 = r_{x/y}^2 \left[ C_{x/y}^2 + K^2 \left( C_{R_1}^2 + C_{R_2}^2 \right) \right]
\]

Thus \( S_{r_1} \) was calculated.

After the test for significant short-term fractionation had been applied to each sub-run and the ratios had been corrected where necessary, each sub-run consisted of one estimate of \( \frac{208 \text{Pb}}{206 \text{Pb}} \) and estimates of several other ratios calculated to be equivalent in time to that estimate.

The next step was to decide whether or not there had been significant fractionation in the course of the run as a whole. Once again
the decision was based upon a comparison of the apparent overall change in $^{208}\text{Pb}/^{206}\text{Pb}$ during the run with the precision of each $^{208}\text{Pb}/^{206}\text{Pb}$ determination. It was not sufficient however, simply to test the difference between successive corrected determinations of $^{208}\text{Pb}/^{206}\text{Pb}$ for significance as all determinations needed to be assessed simultaneously. This was done by a weighted linear regression of $^{208}\text{Pb}/^{206}\text{Pb}$ versus time. The decision regarding the significance of the fractionation which was observed during the run was thus reduced to a decision about the significance of the slope of the regression line.

To help the operator visualise the sample's behaviour more easily, the $^{208}\text{Pb}/^{206}\text{Pb}$ estimate for each sub-run (and its two-standard-error limits) was plotted against time on a 70 x 30 point graph. It was possible at that stage for nominated sub-runs to be removed from consideration in the calculations to follow. This facility was used rarely, as subjective editing may have biased the final results.

A weighted linear regression by least squares was then performed (Guest, 1961). Time was the independent variable and assumed to be without error; weighting of each determination of $^{208}\text{Pb}/^{206}\text{Pb}$ was inversely proportional to the square of the standard error of the mean.

If the line of best fit was to be expressed as

$$f(x) = b_1 x + b_0$$

it was the aim of the least squares method to minimise the sum of the weighted squares of the deviations of each point from the line in the y direction (Q). For $i = 1,2,3, \ldots$ and points $(x_i, y_i)$ with weightings $\omega_i$ where $\omega_i = 1/S_i^2$, and $S_i$ was the standard error of $y_i$, Q was given by

$$Q = \sum \omega_i \{y_i - f(x_i)\}^2$$

The calculations were simplified by making a considered selection of the constant of proportionality between $\omega_i$ and $1/S_i^2$ (P). $P$ was chosen to be $1/\sum(1/S_i^2)$ since
\[ \Sigma \omega_i = \Sigma \left( \frac{P}{S^2} \right) \]
\[ = \frac{P}{\left( \Sigma \frac{1}{S^2} \right)} \]
\[ = \left[ \frac{1}{1} \right] \left[ \Sigma \frac{1}{S^2} \right] \]
\[ = 1 \]

For \( Q \) to be minimised, \( b_0 \) and \( b_1 \) needed to be

\[ b_0 = \frac{\Sigma \omega_i x_i^2 \Sigma \omega_i y_i - \Sigma \omega_i x_i \Sigma \omega_i x_i y_i}{\Sigma \omega_i \Sigma \omega_i x_i^2 - (\Sigma \omega_i x_i)^2} \]

and

\[ b_1 = \frac{\Sigma \omega_i \Sigma \omega_i x_i y_i - \Sigma \omega_i x_i \Sigma \omega_i y_i}{\Sigma \omega_i \Sigma \omega_i x_i^2 - (\Sigma \omega_i x_i)^2} \]

which, since \( \Sigma \omega_i = 1 \), simplified to

\[ b_0 = \frac{\Sigma \omega_i x_i^2 \Sigma \omega_i y_i - \Sigma \omega_i x_i \Sigma \omega_i x_i y_i}{\Sigma \omega_i x_i^2 - (\Sigma \omega_i x_i)^2} \]

and

\[ b_1 = \frac{\Sigma \omega_i x_i y_i - \Sigma \omega_i x_i \Sigma \omega_i y_i}{\Sigma \omega_i x_i^2 - (\Sigma \omega_i x_i)^2} \]

The slope of the line of best fit (\( b_1 \)) having been determined, it was tested for significant deviation from the horizontal. A 't' test was performed on the slope and its variance. The variance of the slope was given by

\[ \text{var } b_1 = \frac{P\Sigma \omega_i}{(\Sigma \omega_i \Sigma \omega_i x_i^2 - (\Sigma \omega_i x_i)^2)} \]

which simplified to

\[ \text{var } b_1 = \frac{P}{(\Sigma \omega_i x_i^2 - (\Sigma \omega_i x_i)^2)} \]
was given by

\[ t = \frac{|b_1|}{\sqrt{\text{var} b_1}} \]

The slope was considered to be significant at the 5% level if 't' equalled or exceeded a fixed value. This value was a function of the number of points in the regression (Acton, 1959) (Table B.1).

<table>
<thead>
<tr>
<th>Degrees of Freedom</th>
<th>Number of Points</th>
<th>t</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>3</td>
<td>6.3</td>
</tr>
<tr>
<td>2</td>
<td>4</td>
<td>2.9</td>
</tr>
<tr>
<td>3</td>
<td>5</td>
<td>2.4</td>
</tr>
<tr>
<td>4</td>
<td>6</td>
<td>2.1</td>
</tr>
<tr>
<td>5</td>
<td>7</td>
<td>2.0</td>
</tr>
</tbody>
</table>

A significance test could not be performed on a line which was defined by two points.

As in the evaluation of sub-runs, the nature of further computation was dependent upon the outcome of the significance test. If the slope was found not to be significant, it was assumed that there had been no fractionation during the run and all determinations in the run were considered to be direct estimates of the various isotopic ratios. All that remained was for replicate determinations of particular isotopic ratios to be identified and for their weighted means to be calculated, in order to arrive at a set of best estimates for the run. The weighted mean and its uncertainty were calculated as outlined previously.

If fractionation was found to have been significant, normalisation was necessary, but because a single Pb spike was used and there was no ratio for which the absolute value was known, the selection of a given \(^{208}\text{Pb}/^{206}\text{Pb}\) was a subjective decision. The graph was a useful aid in the making of this decision for it illustrated changes in the data quality and the fractionation rate through the run. It was left to the operator to decide
at what stage in the run the sample was giving the most reliable data and to feed that time back into the calculations. The most common choice was a time in the middle of the run, on the premise that early emission would be 'light' and that as the beam became heavier with time, its composition would approach more closely, and eventually become 'heavier' than, the true mean sample composition.

The time having been nominated, a best estimate of \(^{208}\text{Pb}/^{206}\text{Pb}\) was calculated from the equation of the regression line. The uncertainty limits for this estimate were dependent upon the particular time selected for they were a function of the uncertainty in the fitted line.

\[
\text{var } f(x) = \sigma \left\{ 1 + (x - \bar{x})^2 \cdot \frac{(\sigma_{\bar{x}})^2}{\sigma_{\bar{x}}^2 \cdot \sigma_{x}^2 - (\sigma_{\bar{x}} \cdot \sigma_{x})^2} \right\}/\sigma_{\bar{x}}
\]

which simplified to

\[
\text{var } f(x) = \sigma \left\{ 1 + (x - \bar{x})^2 \cdot \frac{1}{\sigma_{\bar{x}}^2 \cdot \sigma_{x}^2 - (\sigma_{\bar{x}} \cdot \sigma_{x})^2} \right\}
\]

Differences between the best estimate of \(^{208}\text{Pb}/^{206}\text{Pb}\) and estimates in each of the sub-runs were assumed to be due to mass dependent fractionation so the determinations of the other ratios were normalised to the selected time using a fractionation rate computed from the measured and the calculated \(^{208}\text{Pb}/^{206}\text{Pb}\). Uncertainties in the normalised values were calculated as outlined previously.

At the conclusion of the normalisation procedure, several mutually comparable estimates of different isotopic ratios were available. Replicates of particular ratios were identified and their weighted means and uncertainties calculated to give the best estimates of each for the run.
B.2 THE CORRECTION OF ISOTOPIC ANALYSES FOR FRACTIONATION AND CONTAMINATION

In the isotopic analysis of lead by single-filament, solid-source mass spectrometry, two of the major sources of systematic error are laboratory contamination and variable mass discrimination (or fractionation). Although neither is a problem which is peculiar to the lead system, both are difficulties which lead analysts in particular must cope with.

First, airborne dust is lead-rich (the dust tested at Earth Sciences, ANU contained 600 ppm lead) and fallout into samples must always be minimised. This is particularly so in the case of zircon where small amounts of highly radiogenic lead are measured. Average contamination levels may be determined by the regular analysis of total-processing blanks but, because of the randomness of fallout, these may differ from batch to batch (and probably, from sample to sample) by a factor of 3 or more. Total-processing blanks provide no more than a general estimate of the level of total contamination and a maximum value for the reagent blank.

Secondly, unlike strontium, natural lead contains no two isotopes in fixed relative abundances by which the mass discrimination in individual runs might be corrected. Instead, the mass discrimination induced by a particular mass spectrometer must be established by the repeated analysis of lead of a known isotopic composition under standardised operating conditions and the fractionation of unknown samples be assumed to be the same.

Both these methods of correction are less than satisfactory. Compston and Oversby (1969) showed that a precise correction for mass discrimination was possible if a double spike (one enriched in both $^{207}\text{Pb}$ and $^{204}\text{Pb}$ for instance) was used. Their calculation was an iterative one and several authors have since proposed exact solutions which are more amenable to calculation or more easily visualised than the original (Dallwitz, 1970; Dodson, 1970; Gale, 1970; Hofmann, 1971; Russell, 1971 and others). In none of these solutions however, was the incorporation of random laboratory contamination catered for as well. Gale (1970)
did point out the errors introduced by the neglect of such contamination but concluded that the problem was insoluble given the data normally collected; three double-spiked mixtures were necessary before an exact solution could be obtained.

Knight and Tatsumoto (1972) outlined a means by which corrections for both fractionation and contamination might be made. Their treatment required an estimate of the blank lead in both the spiked and unspiked mixtures to be supplied however, before fractionation factors and the abundances of the sample isotopes could be calculated. Similarly to the solution proposed by Compston and Oversby (1969), the calculation was performed in an iterative manner until successive results differed by less than 0.1 percent.

The problem of the simultaneous correction of lead analyses for fractionation and contamination was reconsidered as part of the present study by Woodford (unpublished manuscript). By extending the vector approach introduced by Russell (1971) to include the blank contribution to both the spiked and unspiked mixtures, Woodford demonstrated that, for the common case in which the relative abundances of four lead isotopes in the spiked and unspiked mixtures and the blank are known, in addition to the absolute abundances of four isotopes in the spike, the system is under-determined and an exact solution is not possible. On the other hand, by a suitable assumption a one-parameter solution may be obtained. If the value of one variable (for instance, the blank contribution to the unspiked mixture) can be estimated, all the other variables (for instance, the blank contribution to the spiked mixture, the absolute abundances of the sample isotopes and mass discrimination factors) are uniquely determined.

Experimentation with the use of Woodford's method to reduce the single-spiked zircon analyses which were done as part of the present study showed that the corrections for fractionation were considerably smaller than those for contamination. For this reason, it was decided that
fractionation was best treated as being zero, in which case two variables were eliminated and the system was no longer underdetermined but overdetermined.

B.2.1 An outline of the mathematical solution

Let a solution containing sample lead of an unknown composition be split into two portions in a known proportion. To one portion a known mass of spike of known composition is added. In the course of processing, let unknown amounts of lead contamination of known composition be introduced to both portions.

Let $S$ be the 4-vector of isotope amounts in picomoles for the sample lead in the spiked mixture.

$k$ be the ratio of the number of moles of sample left unspiked to the number of moles of sample spiked (hence $kS$ is the 4-vector of sample isotope amounts in the unspiked mixture).

$T$ be the 4-vector of isotope amounts in picomoles for the spike added to the spiked mixture.

Let the order of isotopes in the components of these 4-vectors be $1 - ^{206}\text{Pb}, 2 - ^{204}\text{Pb}, 3 - ^{207}\text{Pb}, 4 - ^{208}\text{Pb}$.

Let $B$, $M$ and $M$ be the 4-vectors of isotope ratios with respect to $^{206}\text{Pb}$ for the blank, the measured ratios in the spiked mixture and the measured ratios in the unspiked mixture respectively.

Let $b_s$ and $b_u$ be the amounts in picomoles of $^{206}\text{Pb}$ in the blank components of the spiked and unspiked mixtures respectively.

Let $m_s$ and $m_u$ be the total amounts in picomoles of $^{206}\text{Pb}$ in the spiked and unspiked mixtures respectively. Mass balance equations may be written as follows:

For the spiked mixture,

$$S + b \overset{\text{sp}}{B} + T = m \overset{\text{sp}}{M}$$

(1)
For the unspiked mixture,

\[ kS + bB = mM \]  \hspace{1cm} (2)

These are eight equations in eight unknowns \( S_1, S_2, S_3, S_4, b_u, b_s, m_s \) and \( m_u \) but the equations are not linearly independent. Elimination of the vector \( S \) from the equation gives

\[ \frac{mM}{S} - \frac{m_u}{k} - \frac{b_u}{k} - b_s = T \]  \hspace{1cm} (3)

Note that the two unknowns representing the amount of blank have coalesced to form a single unknown variable:

\[ X_3 = \frac{b_u}{k} - b_s \]  \hspace{1cm} (4)

The implication of equation (4) is that the most information obtainable about the blank components is the value of \( X_3 \) and that they are related by the linear expression (4). If \( X_3 \) can be obtained, the estimation of either \( b_u \) or \( b_s \) enables the other to be calculated.

If new variables are defined

\[ X_1 = m_s \quad \text{and} \quad X_2 = -\frac{m_u}{k} \]

the following system of linear equations follows from equation (3)

\[ X_1 M + X_2 M + X_3 B = T \]  \hspace{1cm} (5)

This represents four equations in three unknowns \( X_1, X_2 \) and \( X_3 \) which can be solved by the method of least squares. From the original mass balance equation (1) the number of picomoles of sample lead in the spiked mixture can be obtained as

\[ S = X_1 M_s - T - b_s B \]  \hspace{1cm} (6)

and the amount of sample lead in the unspiked mixture is given by the
vector $kS$.

The possible values of $b_s$ and $b_u$ are constrained by physical limitations. The amount of blank cannot be negative so the lower bound is fixed

$$b_s \geq b_s^{\text{min}} = \max (X_3, 0)$$

which in turn implies the lower bound

$$b_u \geq b_u^{\text{min}} = \max (-kX_3, 0)$$

From the mass balance equations (1) and (2) an upper bound for the blank components is obtained since no isotope may have a negative abundance. Therefore,

$$b_s \leq b_s^{\text{max}} = \min_{i=1,2,3,4} \left\{ \frac{(m_i M - T)}{S_i T} \right\}$$

and from equation (4)

$$b_u^{\text{max}} = b_s^{\text{max}} + kX_3$$

The maximum allowable range in the blank in the unspiked mixture is consequently

$$b_s^{\text{min}} \leq b_s \leq b_s^{\text{max}}$$

This may be used as a guide when selecting a value for $b_s$ for insertion into equation (6) in order to obtain the vector $S$.
APPENDIX C

ANALYTICAL PRECISION AND ACCURACY

C.1 CALIBRATION OF THE MSZ MASS SPECTROMETER

C.1.1 Electrometer response

When isotopic ratios were determined by rapid peak switching, a correction for the finite response time of the electrometer was necessary. The Cary Model 36 Electrometer which was fitted to the MSZ mass spectrometer was designed specifically for the rapid measurement of ion currents but, with a $5 \times 10^{10}$ ohm input resistor, its response time was significant in comparison to the delay of one second which was allowed for the switching of peaks. The correction was calculated from an estimate of the 'dynamic zero'.

The dynamic zero was the difference between estimates of the background obtained first by static counting and secondly by rapid peak switching between the background and a large reference signal. This simulated the determination of an isotopic ratio and provided a measure of the effect of electrometer memory. The magnitude of the dynamic zero was closely related to the magnitude of the reference signal. It was established empirically that the dynamic zero for an integration time of one second and a one-second delay was of the order of 0.015% if the reference signal was equivalent to full scale deflection on the one volt range.

In the analysis of rubidium and strontium, where isotopic ratios rarely exceeded 10:1, an accurate determination of the dynamic zero was not necessary. It was sufficient, at the commencement of each mass spectrometer run, to switch between the largest of the sample peaks and the background twenty to thirtytimes and to calculate the proportion of the reference peak which the dynamic zero represented. This constant proportion was used in the correction of isotopic ratios for the remainder of the run.
In the analysis of highly radiogenic samples (for instance, lead from zircon or monazite) this procedure was not satisfactory. An accurate estimate of isotopic ratios of the order $10^3$ to $10^4$ was required and the contribution of electrometer memory to the smaller peak was a very significant one. It was necessary to determine the dynamic zero with as great an accuracy as were the isotopic ratios themselves and to test the validity of the earlier assumption that the relationship between the size of the dynamic zero and the size of the reference signal was linear.

Three runs were done in which the specific intention was to define the dynamic zero. They were spaced out over a period of four days. A strontium sample was used as a source of a stable reference beam. In the course of the runs, the beam intensity was adjusted in steps between extremes of about $2 \times 10^{-12}$ amps and $3 \times 10^{-10}$ amps so that the dynamic zero was calculated for reference signals from 100 millivolts to 10 volts.

At each step, an estimate of the dynamic zero was obtained by switching thirty times between the background position and the reference signal, then correcting for the true background using the mean of sixty estimates of the static zero, thirty made before and thirty made after the peak switching. When the static zero was stable throughout a run or a steady drift with time was obvious, all the estimates for the run were combined in the calculation of the background correction for data acquired on each electrometer range.

From the three runs, a total of one hundred and twenty three mean estimates of the dynamic zero, obtained on ranges of one volt, two volts, five volts and ten volts and with delays after switching of one to five seconds, were calculated. The data provided an answer to several questions.

1) Was the dynamic zero a simple function of the beam size and, if so, was the relationship linear?
2) Was the dynamic zero sensitive to the electrometer range used?
3) In what way was the dynamic zero dependent upon the delay time after
switching and was it practicable to increase the delay to the extent that the dynamic correction became insignificant?

4) Was the relationship between the dynamic zero and the size of the reference beam a machine constant?

Figure C.1A illustrates the manner in which the dynamic zero was related to the size of the reference peak. So that a comparison could be made between the data acquired on different ranges, the counts observed were multiplied by the full scale deflection voltage for the range on which each measurement was made. The data plotted are from a single run during which a delay and an integration time of one second was used. The relationship was non-linear and once this was known, it was evident that isotopic ratios would have to be corrected using a dynamic zero which was appropriate to the size of the beam which was being measured.

The figure also illustrates the lack of dependence of the dynamic zero on the electrometer range. For a particular reference signal, estimates of the dynamic zero which were obtained on different ranges were not systematically different from one another.

Figure C.1B shows the effect of increasing the delay time to as much as five seconds. Although the data were sparse, the non-linearity of the relationship between the size of the reference peak and the dynamic zero appeared to be maintained independently of the delay time. From the curves it was calculated that the memory in the electrometer could be modelled by the decay of an RC circuit which had a time constant of approximately 3.5 seconds. In that case, a delay of considerably more than five seconds would have been necessary before the dynamic correction was reduced to insignificance. As this would have nullified the advantages of rapid peak switching, it was not adopted.

A long term instability in the dynamic zero was found during the period of the experiment. This can be seen in the difference between the curves which correspond to delay and integration times of one second in Figs. C.1A and C.1B, the data for which were collected two days apart. No
Fig. C.1A The relationship between the dynamic zero and the size of the reference peak, measured June 1st, 1975. Over the entire range of reference signals the relationship was non-linear and independent of the electrometer range on which the measurements were made.

Fig. C.1B The dependence of dynamic zero upon delay time, measured May 29th, 1975. The non-linearity of the relationship between the dynamic zero and the size of the reference signal persisted for increased delay times. The difference between the one-second delay, one-second integration curve and that in Fig. C.1A is evidence for long term instability in the dynamic zero over the two days which separated the experiments.
FIGURE C.1
satisfactory explanation for the instability was found (with the possible exception of temperature fluctuations having modified the RC characteristics of the electrometer circuitry) and the effect was not observed when the experiment was repeated several months later.

Subsequently to the work described above, paper dielectric capacitors in the electrometer's output filter were replaced by capacitors with a polystyrene dielectric. The intention was to decrease the response time and to improve the linearity. The change had the desired effect. The dynamic zero became a linear function of the size of the reference peak for signals up to about 700,000 (counts/sec) volts and it was reduced by approximately a factor of four.

In Fig. C.2A, the dynamic zero for delay and integration times of one second (the conditions which were appropriate to the routine determination of isotopic ratios) is shown as it was measured soon after the change. Comparable data collected more than twelve months later are illustrated in Fig. C.2B. In both experiments, non-linearity was observed for reference signals greater than approximately 700,000 (counts/sec) volts but the response was linear up to this beam intensity. The change in the nature of the non-linearity over the twelve month period was not readily attributable to any single factor for several worn components in the main amplifier of the electrometer were replaced in that time. As no change in the dynamic zero for reference signals less than 700,000 (counts/sec) volts was observed, it was assumed that the dynamic correction of 0.0042% which had been applied to all the lead and uranium data (the great majority of which were collected during 1976) was a valid one.

In view of the increase in the speed of response of the electrometer which was a consequence of the replacement of the paper dielectric capacitors, the question of a significant reduction in the dynamic correction (or even its elimination) being achieved by an increase in the delay time was re-examined. The dynamic zero for an integration time of one second and delay times from 0.1 to 5.0 seconds was determined. A reference signal of
Fig. C.2A  The relationship between the dynamic zero and the size of the reference signal, measured October 7th, 1975. Non-linearity is evident only for signals exceeding 700,000 (counts/sec) volts.

Fig. C.2B  The relationship between the dynamic zero and the size of the reference signal, measured October 31st, 1976. A curve approximating the data plotted in Fig. C.2A has been drawn for comparison. As in that data, non-linearity is evident only above 700,000 (counts/sec) volts.
FIGURE C.2
650,000 (counts/sec) volts was selected for the experiment as it was the largest signal for which the relationship between the dynamic zero and the reference signal was assuredly linear. The data collected are illustrated in Fig. C.3A. The two-standard-error limits shown were calculated from replicate determinations.

The significant feature of the data was the 'shoulder' in the defined curve at a delay time of about 0.9 seconds. Since the integration time was one second, the implication was that the residual signal did not decay exponentially with time, but rather passed through a local minimum in the time interval zero to one second.

An approximation to the variation in the strength of the residual signal with time is plotted in Fig. C.3B. The curve was calculated by assuming that, for delay times in excess of two seconds, the decay was exponential with time. On that basis, the mean signal size at \( 2.5 \pm 0.1 \) seconds was calculated to be approximately a factor of \( 3 \times 10^{-5} \) the size of the reference signal. The period from 0.1 to 5.5 seconds was then subdivided into intervals of 0.1 seconds duration and, using the mean signal at \( 2.5 \pm 0.1 \) seconds as a reference point, a smooth curve which satisfied the one second integrations of dynamic zero which had been measured (Fig. C.3A) was determined empirically. For simplicity, the calculation was done using \( 0.5 \times 10^{-5} \) times the reference peak as the smallest unit of signal size.

The curve defined by the data in Fig. C.3B reflected the combination of at least two effects, the exponential decay of one RC circuit which had a time constant of about 1.5 seconds and the underdamped decay of a feedback circuit (that was also an RC circuit) which had a time constant less than 0.15 sec.

The experiment showed that even after a delay of five seconds, the dynamic zero remained distinguishable from the background noise and that, if measurements were to be done by rapid peak switching, a significant correction for dynamic effects was unavoidable.
Fig. C.3A Dynamic zero, expressed as a proportion of the reference peak, plotted against the delay time between switching and the commencement of the one-second integration. Two standard error limits are shown where they exceed the size of the point. The inflexion in the curve at about 0.9 seconds was not consistent with simple exponential decay of the residual signal.

Fig. C.3B The decay of the residual signal as inferred from Fig. C.3A. The calculation was simplified by predetermining the signal size after 2.5 seconds. The pattern was explained by a combination of the effects of the underdamped oscillation of a feedback circuit combined with the decay of an RC circuit with a much longer time constant.
FIGURE C.3
On the other hand, it also showed that there may have been a significant advantage in increasing the delay time to 1.5 seconds. First, a reduction of nearly 25 percent in the dynamic correction would have been achieved, secondly, the tail of the underdamped oscillation would have been avoided and, thirdly, the integration would not have been commenced at a time when the signal strength was changing very rapidly. A limited number of experiments which were done using a 1.5 second delay however, showed no significant increase in precision so, for the sake of internal consistency, a delay time of one second was used for the remainder of the analyses which were done as part of the present study.

C.1.2 The measurement of small signals

Because the Cary 36 which was fitted to the MSZ mass spectrometer was not an auto-ranging electrometer, the problem of measuring very large ratios was, in essence, a problem of measuring very small signals. There were three principal factors which limited the accuracy of such measurements: the response time of the electrometer, the zero-point noise and the linearity of the detection and measurement system. The correction for electrometer response could be done very accurately using the data collected in the experiments described in Section C.1.1 and the effects of zero-point noise could be minimised by taking repeated measurements on the largest possible signals but only a limited amount of information was available about linearity.

On the basis of an earlier study of linearity (Compston et al., 1972) it was decided that the component in the detection and measurement system which was most likely to behave non-linearly at the extremes of its operating range was the voltage to frequency converter. By feeding accurately known voltages into the feedback loop of the electrometer, it was possible to test the entire measurement system, with the exception of the input resistor.

The voltages were supplied by a mercury cell and metered by a
voltage divider and variable resistor. They were monitored continuously using a Hewlett Packard 3465A Digital Multimeter with a specified accuracy of 0.02\% of the range plus 0.03\% of the reading. In that way it was ensured that the required voltage at all times was supplied.

The input voltage was varied in a stepwise fashion between 0.1mV and 1V. All measurements were made on the electrometer's 1 Volt range. The data collected consisted of the means of sets of twenty observations, corrected for background by the means of twenty estimates of the zero point which were made before and after each set. Integration times up to nine seconds were used, for in some cases the signal was so small that the problem of detecting it above the zero-point noise became acute. Up to sixteen means of sets of measurements done in this way were combined to give a single estimate of the output frequency which corresponded to a particular input voltage. The standard error of the best estimate was calculated from the replicates. The results of the experiment are summarised in Table C.1 and plotted in Fig. C.4. The frequency measured for an input signal of 1V was used as a reference and the frequency expected for smaller input voltages on that basis was compared with that which was observed. The parameter $\Delta$ which is plotted in Fig. C.4 was calculated as

$$\Delta = \left( \frac{\text{Frequency observed}}{\text{Frequency expected}} - 1 \right) \times 10^2$$

Given the analytical uncertainties, no systematic deviation of $\Delta$ from zero could be detected for any input voltage. It was therefore concluded that non-linearity of the spectrometer's measurement system was unlikely to be a major source of systematic error in determinations of very large ratios.
TABLE C.1
LINEARITY CALIBRATIONS

<table>
<thead>
<tr>
<th>Input Voltage</th>
<th>Counts/Sec expected*</th>
<th>Counts/Sec observed*</th>
<th>Δ φ</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0V</td>
<td>101533.7 ± 50.8</td>
<td>101533.7 ± 2.4</td>
<td></td>
</tr>
<tr>
<td>1.00mV</td>
<td>101.533 ± .051</td>
<td>101.55 ± .24</td>
<td>.017 ± .242</td>
</tr>
<tr>
<td>0.90mV</td>
<td>91.380 ± .046</td>
<td>91.04 ± .46</td>
<td>- .372 ± .503</td>
</tr>
<tr>
<td>0.50mV</td>
<td>50.767 ± .025</td>
<td>50.89 ± .36</td>
<td>.242 ± .709</td>
</tr>
<tr>
<td>0.20mV</td>
<td>20.307 ± .010</td>
<td>19.83 ± .36</td>
<td>-2.349 ± 1.773</td>
</tr>
<tr>
<td>0.18mV</td>
<td>18.276 ± .009</td>
<td>18.39 ± .52</td>
<td>.624 ± 2.845</td>
</tr>
<tr>
<td>0.10mV</td>
<td>10.153 ± .005</td>
<td>9.96 ± .36</td>
<td>-1.901 ± 3.546</td>
</tr>
<tr>
<td>0.08mV</td>
<td>8.123 ± .004</td>
<td>8.34 ± .22</td>
<td>2.677 ± 2.709</td>
</tr>
<tr>
<td>0.05mV</td>
<td>5.077 ± .003</td>
<td>5.15 ± .27</td>
<td>1.444 ± 5.318</td>
</tr>
<tr>
<td>0.04mV</td>
<td>4.061 ± .002</td>
<td>3.75 ± .44</td>
<td>-7.665 ± 10.834</td>
</tr>
<tr>
<td>0.03mV</td>
<td>3.046 ± .002</td>
<td>3.24 ± .13</td>
<td>6.369 ± 4.268</td>
</tr>
<tr>
<td>0.02mV</td>
<td>2.031 ± .001</td>
<td>1.87 ± .66</td>
<td>-7.914 ± 32.501</td>
</tr>
<tr>
<td>0.01mV</td>
<td>1.015 ± .001</td>
<td>1.27 ± .28</td>
<td>25.086 ± 27.578</td>
</tr>
</tbody>
</table>

* Uncertainties expressed as two standard errors of the mean.

+ Uncertainty calculated from the specified accuracy of the digital voltmeter.

§ Calculated uncertainty two standard errors, approximating the uncertainty specified for the digital voltmeter as being two standard errors.
Fig. C.4 The relationship between $\Delta$ and input voltage established in June 1976 for the MSZ spectrometer. The measurement system fitted to the spectrometer at the time included a Cary Model 36 electrometer (#74) and a Hewlett Packard 12501 A voltage to frequency converter (#730-00414). Uncertainty estimates are two standard errors of the mean.
\[ \Delta = \left( \frac{\text{Frequency Observed}}{\text{Frequency Expected}} - 1 \right) \times 10^2 \]
C.2 ANALYSES OF STANDARDS

C.2.1 Rubidium-strontium standards

C.2.1.1 SRM987 strontium standard

Analyses of the NBS standard strontium SRM987 which were done at Earth Sciences, A.N.U. on the mass spectrometer used in the present study have been summarised by Compston et al. (1977b). As was the procedure for all strontium analyses, the measured ratios were corrected by linear normalisation to the standard value for $^{88}\text{Sr}/^{86}\text{Sr}$ of 8.3752, which in most cases resulted in a correction of between 0.2 and 0.3% per mass unit.

The mean normalised $^{87}\text{Sr}/^{86}\text{Sr}$ for twenty-nine different runs of unspiked SRM987 strontium which were done by several analysts over the period from early 1973 to mid 1976 was 0.71029, with a standard error for a single observation of 0.00006. The mean normalised $^{87}\text{Sr}/^{86}\text{Sr}$ for eleven runs of SRM987 spiked with a mixed $^{85}\text{Rb}-^{84}\text{Sr}$ tracer which were done after early 1975 was 0.71031 and the population standard error was 0.00003. Compston et al. (1977b) concluded that no significant bias could be detected between their measurements and those of Nyquist et al. (1974).

C.2.1.2 SRM607 K-feldspar standard

Analyses of the standard potassium feldspar SRM607 for which the '85127' mixed $^{85}\text{Rb}-^{84}\text{Sr}$ spike was used have been summarised by Williams et al. (1976). From six runs, a mean $^{87}\text{Rb}/^{86}\text{Sr}$ of 24.180 ± 0.004 (s.e.) and a mean normalised $^{87}\text{Sr}/^{86}\text{Sr}$ of 1.20114 ± 0.00015 (s.e.) were calculated.

Replicate analyses of the same standard, but using the $^{85}\text{Rb}-^{84}\text{Sr}$ spike '746895' (which superseded '85127'), are listed in Table C.2. The mean $^{87}\text{Rb}/^{86}\text{Sr}$ of 24.167 ± 0.037 (s.e.) and the mean normalised $^{87}\text{Sr}/^{86}\text{Sr}$ of 1.20061 ± 0.00023 (s.e.) are equivalent to an age of 1415.3 ± 2.2 (s.e.) m.y. if an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7100 is assumed. The mean of the ages which were calculated from each run is 1415.3 ±
The standard error of the age is independent of whichever of the two methods of calculation is used so no significant covariance between $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ can be detected. The analysis of SRM607 reported by the U.S. National Bureau of Standards is equivalent to an age of 1412.0 m.y.

### Table C.2

Replicate rubidium and strontium analyses of SRM607 standard K-feldspar using the '746895' *mixed Rb-Sr tracer*

<table>
<thead>
<tr>
<th>Operator</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><em>$^{87}\text{Sr}/^{87}\text{Rb}$</em></th>
<th>Age (.7100)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MM</td>
<td>520.6</td>
<td>65.09</td>
<td>24.211</td>
<td>1.20102</td>
<td>0.02028</td>
<td>1413.9</td>
</tr>
<tr>
<td>MM</td>
<td>519.7</td>
<td>65.28</td>
<td>24.097</td>
<td>1.20134</td>
<td>0.02039</td>
<td>1421.5</td>
</tr>
<tr>
<td>DJM</td>
<td>527.1</td>
<td>65.81</td>
<td>24.245</td>
<td>1.20141</td>
<td>0.02027</td>
<td>1413.1</td>
</tr>
<tr>
<td>JFF</td>
<td>525.3</td>
<td>66.07</td>
<td>24.066</td>
<td>1.20055</td>
<td>0.02038</td>
<td>1421.0</td>
</tr>
<tr>
<td>JFF</td>
<td>526.2</td>
<td>66.27</td>
<td>24.028</td>
<td>1.20060</td>
<td>0.02042</td>
<td>1423.4</td>
</tr>
<tr>
<td>JFF</td>
<td>521.4</td>
<td>65.35</td>
<td>24.148</td>
<td>1.20040</td>
<td>0.02031</td>
<td>1415.8</td>
</tr>
<tr>
<td>ISW</td>
<td>523.7</td>
<td>65.08</td>
<td>24.354</td>
<td>1.20019</td>
<td>0.02013</td>
<td>1403.4</td>
</tr>
<tr>
<td>ISW</td>
<td>519.5</td>
<td>64.99</td>
<td>24.190</td>
<td>1.19940</td>
<td>0.02023</td>
<td>1410.5</td>
</tr>
</tbody>
</table>

$24.167 \pm .037 \text{ (s.e.m.)} \quad 1.20061 \pm .00023 \text{ (s.e.m.)}$

**C.2.2 Uranium - lead standards**

**C.2.2.1 SRM981 common lead standard**

Although the ion beams which were obtained from the lead standard SRM981 were inferior in size and stability to those obtained from equivalent amounts of lead from monazite or zircon samples, analyses of that standard were the best available means of determining the accuracy of lead isotopic analyses done on the MSZ mass spectrometer. Eleven replicate analyses of SRM981 which were done by two operators over the period late 1975 to mid 1977 are listed in Table C.3. The composition reported by the U.S. National Bureau of Standards is listed for comparison.
Table C.3
Replicate lead analyses of SRM981 common lead standard

<table>
<thead>
<tr>
<th>Run Number</th>
<th>Operator</th>
<th>$^{208}<em>{\text{Pb}}/^{206}</em>{\text{Pb}}$</th>
<th>$^{207}<em>{\text{Pb}}/^{206}</em>{\text{Pb}}$</th>
<th>$^{204}<em>{\text{Pb}}/^{206}</em>{\text{Pb}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>- NBS*</td>
<td>2.1681 ± .0008</td>
<td>0.91464 ± .00033</td>
<td>0.059042 ± .000037</td>
</tr>
<tr>
<td>6065</td>
<td>ISW</td>
<td>2.1643 ± .0003</td>
<td>0.91407 ± .00003</td>
<td>0.059156 ± .000022</td>
</tr>
<tr>
<td>7183</td>
<td>DAN</td>
<td>2.1620 ± .0002</td>
<td>0.91362 ± .00005</td>
<td>0.059159 ± .000033</td>
</tr>
<tr>
<td>7406</td>
<td>ISW</td>
<td>2.1662 ± .0009</td>
<td>0.91450 ± .00002</td>
<td>0.059004 ± .000045</td>
</tr>
<tr>
<td>7407</td>
<td>ISW</td>
<td>2.1635 ± .0002</td>
<td>0.91399 ± .00046</td>
<td>0.059118 ± .000018</td>
</tr>
<tr>
<td>7724</td>
<td>DAN</td>
<td>2.1595 ± .0001</td>
<td>0.91322 ± .00028</td>
<td>0.059249 ± .000075</td>
</tr>
<tr>
<td>8036</td>
<td>DAN</td>
<td>2.1633 ± .0002</td>
<td>0.91369 ± .00055</td>
<td>0.059161 ± .000106</td>
</tr>
<tr>
<td>8040</td>
<td>DAN</td>
<td>2.1618 ± .0001</td>
<td>0.91361 ± .00010</td>
<td>0.059204 ± .000085</td>
</tr>
<tr>
<td>8106</td>
<td>DAN</td>
<td>2.1625 ± .0002</td>
<td>0.91394 ± .00013</td>
<td>0.059258 ± .000144</td>
</tr>
<tr>
<td>8141</td>
<td>ISW</td>
<td>2.1611 ± .0003</td>
<td>0.91372 ± .00044</td>
<td>0.059169 ± .000089</td>
</tr>
<tr>
<td>8382</td>
<td>ISW</td>
<td>2.1611 ± .0002</td>
<td>0.91366 ± .00019</td>
<td>0.059159 ± .000030</td>
</tr>
<tr>
<td>8383</td>
<td>ISW</td>
<td>2.1616 ± .0002</td>
<td>0.91356 ± .00034</td>
<td>0.059146 ± .000006</td>
</tr>
</tbody>
</table>

\[ 2.1624 ± .0005 \quad 0.91378 ± .00010 \quad 0.059162 ± .000020 \]

\[ \text{s.e.}_m \quad \text{s.e.}_m \quad \text{s.e.}_m \]

* Limits of error for the NBS determination are 95% confidence limits in which are included an allowance for all known sources of possible systematic error. The uncertainty quoted for the other determinations is an estimate of precision, two standard errors of the mean ratio measured for the run.

\[ \text{s.e.}_m \]

Standard error of the mean calculated for the population.

Fig. C.5 is a plot of the measured $^{208}_{\text{Pb}}/^{206}_{\text{Pb}}$ and $^{207}_{\text{Pb}}/^{206}_{\text{Pb}}$ ratios and shows that the dominant factor which caused a dispersion in the analyses was mass dependent fractionation. The effects of laboratory contamination and $^{206}_{\text{Pb}}$ measurement error are minimal. The analyses as a group also are displaced from the recommended values of the U.S. National Bureau of Standards along the fractionation locus. This provides an
Fig. C.5 Analyses of the SRM981 common lead standard done between late 1975 and mid 1977. The errors shown are 95% confidence limits for the NBS analysis and precision estimates of two standard errors of the mean for the other data. Runs by different analysts are distinguished by broken and solid lines on the error boxes. The loci for fractionation, $^{206}$Pb measurement error and laboratory contamination of the NBS composition are shown.
estimate of the interlaboratory bias. The mean $^{208}\text{Pb}/^{206}\text{Pb}$ measured 
(2.1624 ± .0005 (s.e.)) is 0.13 ± .03 (s.e.diff) % per mass unit lower than the recommended value, the mean $^{207}\text{Pb}/^{206}\text{Pb}$ measured (0.91378 ± .00010 (s.e.)) is 0.09 ± .02 (s.e.diff) % per mass unit lower than the recommended value and the mean $^{206}\text{Pb}/^{204}\text{Pb}$ measured (16.9027 ± .0057 (s.e.)) is 0.10 ± .05 (s.e.diff) % per mass unit lower than the recommended value. On this evidence only, the A.N.U. lead analyses are systematically lower than those of NBS by 0.10 ± .02 (s.e.) % per mass unit.

C.2.2.2 SRM983 radiogenic lead standard

The relatively few analyses which were done of the NBS radiogenic lead standard SRM983 are listed in Table C.4 and illustrated in Figs. C.6 and C.7.

Table C.4
Replicate lead analyses of SRM983 radiogenic lead standard

<table>
<thead>
<tr>
<th>Run Number</th>
<th>Operator</th>
<th>$^{208}\text{Pb}/^{206}\text{Pb}$</th>
<th>$^{207}\text{Pb}/^{206}\text{Pb}$</th>
<th>$^{204}\text{Pb}/^{206}\text{Pb}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>NBS*</td>
<td>0.013619±.000024</td>
<td>0.071201±.000040</td>
<td>0.000371±.000020</td>
<td></td>
</tr>
<tr>
<td>7195 ISW</td>
<td>0.013769±.000004</td>
<td>0.071225±.000016</td>
<td>0.000371±.000003</td>
<td></td>
</tr>
<tr>
<td>8250 ISW</td>
<td>0.013591±.000004</td>
<td>0.071124±.000028</td>
<td>0.000359±.000002</td>
<td></td>
</tr>
<tr>
<td>8384 ISW</td>
<td>0.013604±.000002</td>
<td>0.071170±.000048</td>
<td>0.000359±.000004</td>
<td></td>
</tr>
<tr>
<td>8385 ISW</td>
<td>0.013597±.000004</td>
<td>0.071116±.000026</td>
<td>0.000360±.000005</td>
<td></td>
</tr>
</tbody>
</table>

* See footnote to Table C.3.

Mean ratios calculated excluding run 7195. Standard error of the mean calculated for the population.

The sensitivity of such analyses to laboratory contamination is evident in the data from run 7195. Approximately 1μg of radiogenic lead
Fig. C.6  Analyses of the SRM983 radiogenic lead standard done between late 1976 and mid 1977. The errors shown are expressed in the same way as in Fig. C.5. The loci of $^{206}$Pb measurement error, fractionation and laboratory contamination are shown for comparison. The lead composition measured in run 7195 was that of a mixture of radiogenic standard lead and common lead contamination.
Analyses of the SRM983 radiogenic lead standard plotted to show the relationship between $^{204}\text{Pb}/^{206}\text{Pb}$ and $^{208}\text{Pb}/^{206}\text{Pb}$. The effect of fractionation and contamination is such that even the $^{204}\text{Pb}/^{206}\text{Pb}$ measured in run 7195 is within the uncertainty limits of the NBS determination. There is no evidence in the data of systematic error in the measurement of the very small $^{204}\text{Pb}/^{206}\text{Pb}$ ratio.
SRM 983
Radiogenic Lead Standard

FIGURE C.7
was used in the analysis yet 0.3ng of contamination would have been sufficient to give the composition observed. The data for run 7195 were excluded when the mean ratios were calculated (Table C.4). It was not possible to distinguish between the effects of $^{206}$Pb error and fractionation using the data, but if the differences between the analyses and the recommended values were assumed to be the result of fractionation, an interlaboratory bias of $0.10 \pm 0.02$ (s.e.m.) % per mass unit as calculated from analyses of SRM981 was sufficient to explain them. The absence of significant differences which cannot be explained by bias is an indication that the principal sources of systematic analytical error in the determination of very large isotopic ratios have been identified and adequately corrected for.

C.2.2.3 SRM613 trace elements in a glass matrix

SRM613 was the standard against which the mixed $^{208}$Pb-$^{235}$U spike that was used for the zircon and monazite analyses was calibrated. It was not possible therefore to measure the concentration of lead and uranium in the standard independently. The composition of the lead in SRM613 was measured and, for the purposes of the spike calibration was assumed to be

\[
\begin{align*}
208\text{Pb} / 206\text{Pb} & = 2.1596 \\
207\text{Pb} / 206\text{Pb} & = 0.90636 \\
206\text{Pb} / 204\text{Pb} & = 17.073
\end{align*}
\]

SRM613 was chosen for the spike calibration because it was a mixed standard. The most important factor in calibrating the mixed spike was to establish the spike's lead to uranium ratio.

Replicate analyses of lead and uranium concentrations in discs of SRM613 cut from various rods have been reported by Barnes et al. (1973). They noted a 'slight trend toward increasing uranium concentration with increasing rod number'. If the ratio of lead to uranium in individual rods is calculated, the total range is 0.75%. Because a mean value of 1.0318 for the lead to uranium ratio was used when
the mixed $^{208}\text{Pb} - ^{235}\text{U}$ tracer was calibrated, the calibration may be systematically in error due to heterogeneity of the standard by as much as 0.38%. The precision of the calibration was much greater (Table C.5).

<table>
<thead>
<tr>
<th>SRM613 aliquot</th>
<th>$^{208}\text{Pb} (\mu\text{g})$</th>
<th>$^{235}\text{U} (\mu\text{g})$</th>
<th>$^{208}\text{Pb}/^{235}\text{U}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>$4.70396 \times 10^{-2}$</td>
<td>$3.53605 \times 10^{-2}$</td>
<td>1.3303</td>
</tr>
<tr>
<td>8</td>
<td>$4.76876 \times 10^{-2}$</td>
<td>$3.58622 \times 10^{-2}$</td>
<td>1.3298</td>
</tr>
<tr>
<td>12</td>
<td>$4.77692 \times 10^{-2}$</td>
<td>$3.59245 \times 10^{-2}$</td>
<td>1.3297</td>
</tr>
</tbody>
</table>

The coefficient of variation of the determination of $^{208}\text{Pb}/^{235}\text{U}$ was 0.024%. The coefficient of variation of the concentration measurements was much greater (0.84% for the concentration of $^{208}\text{Pb}$, for instance). This is due to the fact that the concentration measurements were sensitive to weighing errors and reflects difficulties with the system of spike dispensing which was used at the time the calibration was done.

C.2.2.4 GA834-A zircon standard

A zircon standard was prepared from zircon extracted from the Cooma Syenite sample GA834-A. A trial analysis of the zircon showed that the ages calculated from $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ were apparently in agreement within analytical error (no comprehensive assessment of the analytical error had been possible before that time).
and that the mean zircon age was indistinguishable from that measured by rubidium-strontium total-rock or biotite or by potassium-argon biotite (Tetley et al., in prep.). The GA834-A zircon was therefore suitable for use in the assessment of analytical error in the measurement of lead to uranium ratios in zircons (see section C.3).

138 mg of zircon in the size range greater than 100μ was hand-picked to better than 99.9% purity. An agate mortar and pestle which previously had been washed thoroughly with cold aqua regia and sub-boiling distilled water was used to grind about 10 mg of the sample, the ground sample was discarded and the washing procedure was repeated. The remaining sample was then ground by hand until a grit-free powder was obtained.

The finely-ground powder was emptied onto weighing paper and quartered and recombined several times using a second piece of weighing paper. Twenty fractions of the powder were then prepared by quartering, each of which was about 5 mg in weight. These were stored in separate sample phials.

Five of the aliquots of powder were selected at random and analysed for uranium and lead using routine procedures. The data are listed in Table C.6 and plotted on a concordia diagram in Fig. C.8. The mean \(^{206}\text{Pb} / ^{238}\text{U}\) is 0.027231 ± 0.000136 (s.e.m) and the mean \(^{207}\text{Pb} / ^{235}\text{U}\) is 0.18662 ± 0.00110 (s.e.m). The coefficients of variation of the population for \(^{206}\text{Pb} / ^{238}\text{U}\) and \(^{207}\text{Pb} / ^{235}\text{U}\) are 1.12 and 1.32% respectively. The significance of the GA834-A standard data is discussed in section C.3.
Table C.6  Uranium-lead analytical data from the Cooma Syenite (GA834-A)
The estimates of precision were calculated assuming the
coefficient of variation for the determination of uranium
relative to lead was 0.5% (see section C.3.2.). The
composition of common lead was calculated by the linear
Model III of Cumming & Richards (1975), assuming an age
of 176 m.y. (Appendix E).
### TABLE C.6
ANALYTICAL DATA FOR THE COOMA SYENITE (GA834-A)

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Sample Weight mg</th>
<th>Pb ppm</th>
<th>U ppm</th>
<th>206(_{\text{Pb}})(^*) measured</th>
<th>ATOM PERCENT RADIOGENIC LEAD</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>206(_{\text{Pb}})</td>
<td>207(_{\text{Pb}})</td>
</tr>
<tr>
<td>88/ # 13</td>
<td>3.97</td>
<td>12.46</td>
<td>387.74</td>
<td>3139±230</td>
<td>73.63</td>
</tr>
<tr>
<td>87/ # 12</td>
<td>6.16</td>
<td>12.72</td>
<td>390.95</td>
<td>3122±239</td>
<td>73.65</td>
</tr>
<tr>
<td>90/ # 20</td>
<td>5.15</td>
<td>12.68</td>
<td>394.02</td>
<td>2929±145</td>
<td>73.69</td>
</tr>
<tr>
<td>89/ # 18</td>
<td>5.79</td>
<td>12.39</td>
<td>394.60</td>
<td>3942±111</td>
<td>73.67</td>
</tr>
<tr>
<td>86/ # 5</td>
<td>5.45</td>
<td>12.74</td>
<td>397.62</td>
<td>3092±141</td>
<td>73.80</td>
</tr>
</tbody>
</table>

*ATOMIC RATIOS

\[
\frac{206\text{Pb}}{238\text{U}}, \quad \frac{207\text{Pb}}{235\text{U}}, \quad \frac{207\text{Pb}}{206\text{Pb}}
\]

<table>
<thead>
<tr>
<th>APPARENT AGES (m.y.)</th>
<th>206(_{\text{Pb}})</th>
<th>207(_{\text{Pb}})</th>
<th>207(_{\text{Pb}})</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>238(_{\text{U}})</td>
<td>235(_{\text{U}})</td>
<td>206(_{\text{Pb}})</td>
</tr>
<tr>
<td>173.7</td>
<td>174.3</td>
<td>182.8</td>
<td></td>
</tr>
<tr>
<td>175.3</td>
<td>177.0</td>
<td>200.4</td>
<td></td>
</tr>
<tr>
<td>173.8</td>
<td>173.5</td>
<td>170.4</td>
<td></td>
</tr>
<tr>
<td>170.1</td>
<td>171.8</td>
<td>195.7</td>
<td></td>
</tr>
<tr>
<td>173.2</td>
<td>172.0</td>
<td>156.4</td>
<td></td>
</tr>
</tbody>
</table>

\*Precision estimates expressed as one standard error.
Fig. C.8  A concordia plot of the uranium-lead data from the Cooma Syenite (Table C.6). The estimates of analytical error are two standard errors.
COOMA SYENITE (GA834-A)

FIGURE C.8
C.3 THE ASSESSMENT OF ANALYTICAL ERROR

As described in Appendix B, it was possible to use the estimates of precision calculated from each set of replicate observations of the same isotopic ratio in a single mass spectrometer run to calculate the precision of the mean value of that ratio expected for the run. In practice, it was found commonly that the precision of the mean ratio calculated using the normalised means of each set was less than predicted from the internal precision of each set. This demonstrated the presence of sources of analytical error which affected the between-set replicates more than they affected the within-set replicates. The most important of these was probably error in the estimate of the zero point of the digital measurement system.

During routine runs, an estimate of the zero point was obtained before each ratio was measured and that estimate was used in the calculation of that ratio. For most purposes this was a very satisfactory procedure, as the difference between successive estimates of the zero point was insignificant in comparison to the signal being measured and had a negligible effect on the isotopic ratio calculated from one set to the next. The regular updating of the estimate of the zero point allowed for any long term drift.

For the measurement of large ratios (greater than about 10:1) this procedure was less satisfactory but it was used unless the ratio exceeded about 100:1. For ratios less than about 100:1, the effect of zero-point noise on the determinations was as described above, with the precision estimated from the normalised set means being less than that estimated from the replicates within a set. The former estimate was considered to be the more correct and was used further in the assessment of analytical error. As the ratio to be measured became increasingly large, the correct estimation of the zero point became increasingly critical. In extreme cases, an error of 1 count/sec in the determination of the zero point caused an error in the measured ratio of 25% or more. The procedure adopted in the background correction of very large ratios was to assess all the mean
estimates of the background which had been obtained during the run. If these were the same within error, the mean background was calculated and the one correction factor applied to all the ratio measurements in the run. If the estimates were not the same to within error, either because of a stepwise change in the background level or a slow drift of the level with time, as many of the estimates as could be were combined in the calculation of the background at the time each ratio was measured. The corrected ratios obtained by either method were then processed through the on-line data reduction described in Section B.1 and the final estimate of the precision obtained from the standard error of the mean of the normalised means.

In addition to obtaining precision estimates for the isotopic ratios measured on the mass spectrometer, other sources of error had to be considered: error in the determination of the ratio of lead to uranium and error in the correction for common lead, for instance.

The problem of assessing the analytical error was overcome by a combination of two approaches. First, the combination of errors for the calculation of the composition and concentration of radiogenic lead from isotopic analyses was calculated and secondly, a zircon was selected for which the lead to uranium ratio was expected to be constant (GA834-A) and the error in the measurement of lead to uranium determined empirically.

C.3.1 Error in the isotopic analysis of lead

The following treatment of errors in lead analyses was largely devised by Dr. W. Compston.

C.3.1.1 Error in $^{206}_{\text{Pb}}$ *

Nomenclature:

Let $Cv(N)$ be the coefficient of variation of N

$\sigma(N)$ be the standard error of N

$\text{var}(N)$ be the variance of N

$R^{A/B}$ be the isotopic ratio of isotopes 20A, 20B
sa, sp and m denote sample, spike and spike-sample mixture respectively.

By definition, \( \text{Cv}(N) = \frac{\sigma(N)}{N} \)

from which it follows that

\[ \sigma^2(N) \equiv \text{var } N = N^2 \text{Cv}^2(N) \]

By using the fact that the amount of radiogenic \(^{206}\text{Pb} \) \( (206^*) \) is related to the amounts of \(^{206}\text{Pb} \) in the sample \( (206_{\text{sa}}) \) and present initially \( (206_1) \) by

\[ 206^* = 206_{\text{sa}} - 206_1 \]

and hence that

\[ \text{var } 206^* = \text{var } 206_{\text{sa}} + \text{var } 206_1 \]

it follows from the definition above that

\[ (206^*)^2 \text{Cv}^2(206^*) = (206_{\text{sa}})^2 \text{Cv}^2(206_{\text{sa}}) + (206_1)^2 \text{Cv}^2(206_1) \quad (1) \]

a) **Determination of \( \text{Cv}(206_{\text{sa}}) \)**

Let \( Q = \frac{206_{\text{sa}}}{206_{\text{sp}}} = \frac{R^{8/6}_m - R^{8/6}_{\text{sp}}}{R^{8/6}_{\text{sa}} - R^{8/6}_m} \)

Let the constant of proportionality relating error in \( Q(\Delta Q) \) to error in \( R^{8/6} \)

\[ R^{8/6}_m (\Delta R^{8/6}_m) \text{ be } \frac{\partial Q}{\partial (R^{8/6}_m)} \]

i.e. \( \Delta Q = \frac{\partial Q}{\partial (R^{8/6}_m)} \cdot \Delta R^{8/6}_m \)

Expressing the errors as fractional errors

\[ \frac{\Delta Q}{Q} = \left[ \frac{\partial Q}{\partial (R^{8/6}_m)} \cdot \frac{R^{8/6}_m}{Q} \right] \cdot \frac{\Delta R^{8/6}_m}{R^{8/6}_m} \]

The term \( \frac{\partial Q}{\partial (R^{8/6}_m)} \cdot \frac{R^{8/6}_m}{Q} \) is the error magnification (E Mag)
Considering now one standard error

\[
\frac{\sigma(Q)}{Q} = E \text{ Mag} \cdot \frac{\sigma(R_8/6)}{R_8/6}
\]

that is \(Cv(Q) = E \text{ Mag} \cdot Cv(R_8/6)\)

Since \(Q = \frac{206_{sa}}{206_{sp}}\) and errors in \(206_{sp}\) are systematic it follows that

\[
Cv(206_{sa}) = E \text{ Mag} \cdot Cv(R_8/6)
\] (2)

b) Determination of \(Cv(206_i)\)

Since \(^{204}\text{Pb}\) is entirely nonradiogenic, \(206_i\) is related to \(204_{sa}\) by the ratio of \(^{206}\text{Pb}\) to \(^{204}\text{Pb}\) in the common Pb \((R_{co}^6/4)\).

\[206_i = R_{co}^{6/4} \cdot 204_{sa}\]

hence

\[Cv(206_i) = \sqrt{[Cv^2(R_{co}^{6/4}) + Cv^2(204_{sa})]}\] (3)

Also, since \(204_{sa} = R_{sa}^{4/6} \cdot 206_{sa}\)

then

\[Cv(204_{sa}) = \sqrt{[Cv^2(R_{sa}^{4/6}) + Cv^2(206_{sa})]}\] (4)

Substitution from (4) into (3) gives

\[Cv^2(206_i) = Cv^2(R_{co}^{6/4}) + Cv^2(R_{sa}^{4/6}) + Cv^2(206_{sa})\] (5)

c) Determination of \(Cv(206^*)\)

Substitution from (2) and (5) into (1) gives

\[\frac{(206^*)^2}{Cv^2(206^*)} = \frac{(206_{sa})^2}{Cv^2(206_{sa})} \cdot [E \text{ Mag} \cdot Cv(R_8/6)]^2 + (206_i)^2[Cv(R_{co}^{6/4}) + Cv^2(R_{sa}^{4/6}) + Cv^2(206_{sa})]\]
which, by further substitution from (2), may alternatively be expressed as

\[
Cv^2(206^*) = [E \, \text{Mag.} \, Cv(R_6^{8/6})]^2 \cdot \frac{(206_{sa})^2 + (206_i)^2}{(206^*)^2}
\]

\[
+ \frac{(206_i)^2}{(206^*)^2} \cdot [Cv^2(R_{co}^{6/4}) + Cv^2(R_{sa}^{4/6})]
\]

(6)

Since \( 206^* = 206_{sa} - 206_i \)

and \( 206_i = R_{co}^{6/4} \cdot 204_{sa} \)

and \( 204_{sa} = R_{co}^{4/6} \cdot 206_{sa} \)

it follows that \( 206^* = 206_{sa} (1 - R_{co}^{6/4} \cdot R_{sa}^{4/6}) \)

(7)

and also that \( \frac{206_i}{206^*} = \frac{206_{sa}}{206^*} \cdot R_{co}^{6/4} \cdot R_{sa}^{4/6} \)

so

\[
\frac{206_i}{206^*} = \frac{R_{co}^{6/4} \cdot R_{sa}^{4/6}}{1-R_{co}^{6/4} \cdot R_{sa}^{4/6}}
\]

(8)

also \( \frac{(206_i)^2 + (206_{sa})^2}{(206^*)^2} = \frac{(R_{co}^{6/4} \cdot R_{sa}^{4/6})^2}{(1-R_{co}^{6/4} \cdot R_{sa}^{4/6})^2} + \frac{1}{(1-R_{co}^{6/4} \cdot R_{sa}^{4/6})^2} \)

\[
= \frac{1 + (R_{co}^{6/4} \cdot R_{sa}^{4/6})^2}{(1-R_{co}^{6/4} \cdot R_{sa}^{4/6})^2}
\]

(9)

Substitution from (8) and (9) into (6) gives

\[
Cv^2(206^*) = [E \, \text{Mag.} \, Cv(R_6^{8/6})]^2 \cdot \frac{1 + (R_{co}^{6/4} \cdot R_{sa}^{4/6})^2}{(1-R_{co}^{6/4} \cdot R_{sa}^{4/6})^2}
\]
\[
\frac{(R_{\co}^{6/4} \cdot R_{\sa}^{4/6})^2}{(1-R_{\co}^{6/4} \cdot R_{\sa}^{4/6})^2} \cdot [Cv^2(R_{\co}^{6/4}) + Cv^2(R_{\sa}^{4/6})]
\]

This expression may be simplified since \( R_{\sa}^{4/6} \ll 1 \) in highly radiogenic samples. As a consequence \( (R_{\sa}^{4/6} \cdot R_{\co}^{6/4})^2 \ll 1 \) and
\[
1 - R_{\sa}^{4/6} \cdot R_{\co}^{6/4} \approx 1.
\]

Expression (10) thus becomes

\[
Cv^2(206^*) = [E \text{ Mag} \cdot Cv(R_{\sa}^{8/6})]^2
\]

\[
+ [Cv^2(R_{\co}^{6/4}) + Cv^2(R_{\sa}^{4/6})] \cdot [R_{\co}^{6/4} \cdot R_{\sa}^{4/6}]^2
\]

C.3.1.2 Error in \( ^{207}\text{Pb}^* \)

Similarly to the case for \( ^{206}\text{Pb} \),

\[
^{207}\text{Pb}^* = ^{207}\text{sa} - ^{207}\text{fi}
\]

\[
= R_{\sa}^{7/6} \cdot ^{206}\text{sa} - R_{\co}^{7/4} \cdot ^{204}\text{sa}
\]

\[
= 206_{\sa} \cdot (R_{\sa}^{7/6} - R_{\co}^{7/4} \cdot R_{\sa}^{4/6})
\]

The expression for error in \( ^{207}\text{Pb}^* \) is thus

\[
Cv^2(207^*) = Cv^2(206_{\sa}) + Cv^2(R_{\sa}^{7/6} - R_{\co}^{7/4} \cdot R_{\sa}^{4/6})
\]

Expanding the right hand side of (11)

\[
Cv^2(R_{\sa}^{7/6} - R_{\co}^{7/4} \cdot R_{\sa}^{4/6}) = \frac{\text{var } R_{\sa}^{7/6} + \text{var}(R_{\co}^{7/4} \cdot R_{\sa}^{4/6})}{(R_{\sa}^{7/6} - R_{\co}^{7/4} \cdot R_{\sa}^{4/6})^2}
\]

Also \( \text{var } R_{\sa}^{7/6} = (R_{\sa}^{7/6})^2 \cdot Cv^2(R_{\sa}^{7/6}) \)
and \( \text{var} \left( R^{7/4} \cdot R^{4/6}_{\text{sa}} \right) = \left( R^{7/4} \cdot R^{4/6}_{\text{sa}} \right)^2 \cdot \text{Cv}^2 \left( R^{7/4} \cdot R^{4/6}_{\text{sa}} \right) \)

\[ = \left[ R^{7/4}_{\text{co}} \cdot R^{4/6}_{\text{sa}} \right]^2 \cdot \left[ \text{Cv}^2 \left( R^{7/4}_{\text{co}} \right) + \text{Cv}^2 \left( R^{4/6}_{\text{sa}} \right) \right] \]  

(14)

(12) may thus be expressed more fully through substitution from (13) and (14).

\[
\text{Cv}^2 \left[ R^{7/6}_{\text{sa}} - R^{7/4}_{\text{co}} \cdot R^{4/6}_{\text{sa}} \right] = \frac{1}{\left( R^{7/6}_{\text{co}} - R^{7/4}_{\text{co}} \cdot R^{4/6}_{\text{sa}} \right)^2}
\]

\[
\cdot \left( R^{7/6}_{\text{sa}} \cdot \text{Cv}^2 \left( R^{7/6}_{\text{sa}} \right) + \left[ R^{7/4}_{\text{co}} \cdot R^{4/6}_{\text{sa}} \right]^2 \cdot \left[ \text{Cv}^2 \left( R^{7/4}_{\text{co}} \right) + \text{Cv}^2 \left( R^{4/6}_{\text{sa}} \right) \right] \right)
\]

Substituting from (15) into (11) and using (2) gives

\[
\text{Cv}^2 (207^*) = \left[ \text{E Mag. Cv} \left( R^{8/6}_{\text{m}} \right) \right]^2 + \frac{1}{\left( R^{7/6}_{\text{co}} - R^{7/4}_{\text{co}} \cdot R^{4/6}_{\text{sa}} \right)^2}
\]

\[
\cdot \left( R^{7/6}_{\text{sa}} \cdot \text{Cv}^2 \left( R^{7/6}_{\text{sa}} \right) + \left[ R^{7/4}_{\text{co}} \cdot R^{4/6}_{\text{sa}} \right]^2 \cdot \left[ \text{Cv}^2 \left( R^{7/4}_{\text{co}} \right) + \text{Cv}^2 \left( R^{4/6}_{\text{sa}} \right) \right] \right)
\]

C.3.2 Error in the lead-uranium ratio

The error in the measurement of lead-uranium ratios could be determined in two ways: by the combination of precision estimates for the measurement of uranium and the measurement of lead, or by an evaluation of replicate analyses of samples in which the lead-uranium ratio was thought or known to be constant.
a) **Error assessment using estimates of precision**

The precision of the uranium analyses for any particular sample was readily determined from the precision of the measurement of $^{235}\text{U}/^{238}\text{U}$. If this was calculated as the coefficient of variation of the population of replicate $^{235}\text{U}/^{238}\text{U}$ determinations, rather than from the internal precision of each set, account was taken both of random error and error due to fractionation in the course of the run. Spectrometer bias, which had no bearing on the estimate of precision, was not allowed for.

The total range of coefficients of variation for the zircon and monazite uranium isotopic measurements was 0.02 to 0.66%. The mean was $0.19 \pm 0.14$ (s.e. \text{pop})%. Because the error magnification factor for the sample-spike mixtures which were prepared ranged only between 1.0 and 1.1, this was a close approximation to the coefficient of variation for the determination of uranium concentration.

In view of the relative imprecision of determinations of $^{207}\text{Pb}$ (due in part to their sensitivity to the common lead correction) the uncertainty in $^{207}\text{Pb}/^{235}\text{U}$ was, for many samples, dominated by the uncertainty in $^{207}\text{Pb}$ and the calculation of a mean coefficient of variation for $^{207}\text{Pb}/^{235}\text{U}$ was therefore of little physical significance. On the other hand, the uncertainty in $^{206}\text{Pb}$ was small with respect to that in uranium and was relatively uniform.

The mean coefficient of variation for $^{206}\text{Pb}$ for the zircon and monazite samples was $0.04 \pm 0.04$ (s.e. \text{pop})% and when combined with the mean coefficient of variation for uranium, gave an expected coefficient of variation for the lead-uranium ratio of $0.20 \pm 0.13$ (s.e. \text{pop})%.

b) **Error assessment using replicate analyses**

There were three sets of replicate lead-uranium analyses from which an estimate of the precision of the measurement of the lead-uranium ratio could be obtained: spike calibrations, analyses of monazites (assuming their concordance) and analyses of the Cooma Syenite zircon
standard. The coefficient of variation for the measurement of the lead-
uranium ratio in the calibration of the spike already has been discussed in
section C.2. It was 0.02%. The mean coefficient of variation calculated
from the replicate determinations of $^{206}\text{Pb}/^{238}\text{U}$ for the six monazite
samples was $0.34 \pm 0.16$ (s.e. $\text{pop}$)%$. These were assumed to be independent
estimates of the population variance. The pooled coefficient of variation
was 0.37%.

In contrast to this and to the coefficient of variation expected
from internal precision, the coefficient of variation for the five analyses
of $^{206}\text{Pb}/^{238}\text{U}$ in the Cooma Syenite zircon standard was 1.12%. This was
significantly greater than the expected coefficient of variation (0.20%) at
the 5% level. The difference could be interpreted in two ways. Either the
error estimates based on internal precision alone did not take into account
all the significant sources of analytical error, or homogeneity in
$^{206}\text{Pb}/^{238}\text{U}$ in the standard had not been achieved. Without further analyses
of the standard zircon being done, the question could not be resolved.

It was concluded that the best estimate of the coefficient of
variation for the determination of the lead-uranium ratio was somewhere in
the range between that determined empirically (0.37% and 1.12%) and that
expected from internal precision (0.20 $\pm$ 0.13 (s.e. $\text{pop}$)%). The value chosen
for the coefficient of variation for the determination of uranium relative
to lead was 0.5%.
APPENDIX D

ANALYSIS OF VARIANCE FOR THE RUBIDIUM-STRONTIUM BIOTITE AGES

A nested classification analysis of variance was done for the rubidium-strontium biotite ages from the Berridale Batholith. The method used was that described by Bennett and Franklin (1954). Both the biotite ages measured by Williams et al. (1976) and the additional determinations which are listed in Table D.1 were included in the analysis, but only after it had been verified that the two sets of data could validly be combined.

For all but three of their analyses, Williams et al. (1976) used the mixed $^{85}\text{Rb}^{84}\text{Sr}$ spike '85127'. The three analyses which were the exceptions and all the subsequent analyses were done with the mixed $^{85}\text{Rb}^{84}\text{Sr}$ spike which superseded this, '746895'. If the data were to be pooled in an analysis of variance, it was essential that any systematic error which may have been introduced by differences between the spike calibrations be assessed and an appropriate correction be made.

Replicate analyses of the standard K-feldspar NBS607 for which the '746895' spike was used are listed in Table C.2. The mean of the calculated ages was 1415.3 ± 2.4(s.e.) m.y. The age calculated from the mean $^{87}\text{Rb}/^{86}\text{Sr}$ (24.167 ± 0.037(s.e.)) and the mean $^{87}\text{Sr}/^{86}\text{Sr}$ (1.20061 ± 0.00023(s.e.)) was 1415.3 ± 2.2(s.e.) m.y. The standard error of the former age was not significantly less than the standard error of the latter so the covariance of $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ was not significant.

Williams et al. (1976) reported replicate analyses of NBS607 which were done using the '85127' spike. From the mean values of 24.180 ± 0.004(s.e.) for $^{87}\text{Rb}/^{86}\text{Sr}$ and 1.20114 ± 0.00015(s.e.) for $^{87}\text{Sr}/^{86}\text{Sr}$ an age of 1416 m.y. (for an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7100) was calculated. No uncertainties for the age were quoted but, assuming the absence of covariance, a standard error of 0.5 m.y. may be calculated from the
<table>
<thead>
<tr>
<th>SAMPLE</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>$\frac{87\text{Rb}}{86\text{Sr}}$</th>
<th>$\frac{87\text{Sr}}{86\text{Sr}}$</th>
<th>AGE (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>BUCKLEYS LAKE ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10C</td>
<td>845.3</td>
<td>8.94</td>
<td>324.2</td>
<td>2.6228</td>
<td>415.1</td>
</tr>
<tr>
<td>10D</td>
<td>848.1</td>
<td>9.24</td>
<td>313.5</td>
<td>2.5591</td>
<td>415.0</td>
</tr>
<tr>
<td>110C</td>
<td>850.9</td>
<td>6.80</td>
<td>457.2</td>
<td>3.4260</td>
<td>417.7</td>
</tr>
<tr>
<td><strong>DALGETY GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>640.2</td>
<td>15.59</td>
<td>127.3</td>
<td>1.4622</td>
<td>415.4</td>
</tr>
<tr>
<td>26A</td>
<td>624.9</td>
<td>10.84</td>
<td>184.3</td>
<td>1.8048</td>
<td>417.4</td>
</tr>
<tr>
<td>26B</td>
<td>625.8</td>
<td>10.45</td>
<td>192.3</td>
<td>1.8531</td>
<td>417.7</td>
</tr>
<tr>
<td>96</td>
<td>598.0</td>
<td>6.59</td>
<td>308.6</td>
<td>2.5339</td>
<td>415.2</td>
</tr>
<tr>
<td><strong>FINISTER GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>161</td>
<td>545.5</td>
<td>9.88</td>
<td>176.0</td>
<td>1.7641</td>
<td>421.5</td>
</tr>
<tr>
<td>165</td>
<td>746.2</td>
<td>8.87</td>
<td>283.0</td>
<td>2.3908</td>
<td>417.6</td>
</tr>
<tr>
<td>160</td>
<td>616.1</td>
<td>5.71</td>
<td>379.5</td>
<td>2.9328</td>
<td>411.7</td>
</tr>
<tr>
<td>*163A</td>
<td>636.1</td>
<td>4.73</td>
<td>500.2</td>
<td>3.6618</td>
<td>414.7</td>
</tr>
<tr>
<td>*163B</td>
<td>611.0</td>
<td>4.15</td>
<td>564.3</td>
<td>4.0592</td>
<td>417.0</td>
</tr>
<tr>
<td><strong>MAFFRA ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>21C</td>
<td>1125.2</td>
<td>5.70</td>
<td>848.3</td>
<td>5.7056</td>
<td>413.9</td>
</tr>
<tr>
<td><strong>NUMBLA VALE ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>29</td>
<td>750.4</td>
<td>32.18</td>
<td>70.12</td>
<td>1.1299</td>
<td>418.5</td>
</tr>
<tr>
<td>2C</td>
<td>1112.4</td>
<td>9.76</td>
<td>407.7</td>
<td>3.1444</td>
<td>418.9</td>
</tr>
<tr>
<td><strong>TARA GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>75</td>
<td>506.8</td>
<td>10.07</td>
<td>158.5</td>
<td>1.6337</td>
<td>410.9</td>
</tr>
<tr>
<td>15</td>
<td>978.8</td>
<td>8.08</td>
<td>437.9</td>
<td>3.2900</td>
<td>414.3</td>
</tr>
</tbody>
</table>

*Samples analysed by W. Compston.  
A, B, C etc. - replicate analyses.  
All samples analysed with 746895 Spike.
uncertainties in $^{87}$Rb/$^{86}$Sr and $^{87}$Sr/$^{86}$Sr. Williams et al. (1976) noted that the difference between 1416 ± 0.5 m.y. and that determined by the U.S. National Bureau of Standards (1412.0 m.y.) was of the same order as an estimate of their random error.

Because the difference between 1416 ± 0.5(s.e.m) m.y. and 1415.3 ± 2.2(s.e.m) m.y. was not significant, the biotite ages measured using each of the spikes were considered to be mutually comparable and no compensation for systematic errors was required.

In the analysis of variance, the data were grouped into three classes.

1) Replicate observations of the age of one sample.
2) Age determinations on samples from the one pluton.
3) Age determinations on plutons.

By doing this it was possible to obtain an estimate of the standard error of a single observation and also to test the significance of the differences between the measured ages of samples from any one pluton and between the measured ages of the plutons themselves. The analysis of variance was done as follows:

Let $T_{ij}$ be the sum of the n observations of the age of any one sample.

$T_i$ be the sum of the nq observations of the age of any one pluton.

$T$ be the sum of all observations, of which there are $N$.

$p$ be the number of plutons, that is, $N = npq$.

$S_i$ be the sum of squares of the deviations of the p pluton means from the overall mean.

$S_{j(i)}$ be the sum of squares of the deviations of the pq sample means from the p pluton means.

$S_{\alpha(ij)}$ be the sum of squares of the deviations of the $N$
observations from the \( pq \) means \( \bar{x}_{ij} \) of the samples, 

\( \sigma^2 \) be the within-sample mean square. 

\( \sigma^2_\lambda \) be the between-samples mean square. 

\( \sigma^2_\varepsilon \) be the between-plutons mean square.

Bennett and Franklin (1954) showed that

\[
S_i = \frac{\sum_i T_i^2}{nq} - \frac{T^2}{N}
\]

and has \( p-1 \) degrees of freedom,

\[
S_{j(i)} = \frac{\sum_{ij} T_{ij}^2 \Sigma_i T_i^2}{n} - \frac{\Sigma_i T_i^2}{nq}
\]

and has \( p(q-1) \) degrees of freedom and that

\[
S_{\alpha(ij)} = \sum_{ij} \alpha{x}_{ij}^2 - \frac{\sum_{ij} T_{ij}^2}{n}
\]

and has \( N-pq \) degrees of freedom.

If the possible number of samples per pluton was assumed to be infinite then

\[
S_i = \sigma^2 + n\sigma^2_\lambda + nq\sigma^2_\varepsilon
\]

and

\[
S_{j(i)} = \sigma^2 + n\sigma^2_\lambda
\]

and

\[
S_{\alpha(ij)} = \sigma^2
\]
The computations leading to the analysis of variance in Table D.3 are shown in Table D.2. To simplify the calculations, 400 m.y. was deducted from all the measured ages.

Table D.2

<table>
<thead>
<tr>
<th>Pluton</th>
<th>Sample Number</th>
<th>Observation</th>
<th>Sample Total</th>
<th>Pluton Total</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>1 2 3 4</td>
<td>T_1j</td>
<td>T_1</td>
</tr>
<tr>
<td>Bimbimbie</td>
<td>74</td>
<td>14.8</td>
<td>14.8</td>
<td>14.8</td>
</tr>
<tr>
<td></td>
<td>22</td>
<td>16.8 17.1</td>
<td>33.9</td>
<td></td>
</tr>
<tr>
<td></td>
<td>33</td>
<td>19.0</td>
<td>19.0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>10</td>
<td>19.1 18.5 15.1 15.0</td>
<td>67.7</td>
<td></td>
</tr>
<tr>
<td>Buckleys Lake</td>
<td>30</td>
<td>16.5</td>
<td>16.5</td>
<td>225.4</td>
</tr>
<tr>
<td></td>
<td>7</td>
<td>22.7 16.5</td>
<td>39.2</td>
<td></td>
</tr>
<tr>
<td></td>
<td>110*</td>
<td>9.3 16.6 17.7</td>
<td>34.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td>109</td>
<td>14.8</td>
<td>14.8</td>
<td></td>
</tr>
<tr>
<td>Cootralantra</td>
<td>36</td>
<td>10.7 13.3</td>
<td>18.0</td>
<td>60.1</td>
</tr>
<tr>
<td></td>
<td>83</td>
<td>18.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>11.3 10.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>19</td>
<td>22.9 19.2</td>
<td>42.1</td>
<td></td>
</tr>
<tr>
<td>Currowong</td>
<td>48</td>
<td>15.3</td>
<td>15.3</td>
<td>15.3</td>
</tr>
<tr>
<td></td>
<td>90</td>
<td>19.5 18.5</td>
<td>38.0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>9</td>
<td>15.4</td>
<td>15.4</td>
<td></td>
</tr>
<tr>
<td>Dalgety</td>
<td>26</td>
<td>17.4 17.7</td>
<td>35.1</td>
<td>133.2</td>
</tr>
<tr>
<td></td>
<td>94</td>
<td>13.1</td>
<td>13.1</td>
<td></td>
</tr>
<tr>
<td></td>
<td>68</td>
<td>16.4</td>
<td>16.4</td>
<td></td>
</tr>
<tr>
<td></td>
<td>96</td>
<td>15.2</td>
<td>15.2</td>
<td></td>
</tr>
<tr>
<td>Delegate</td>
<td>46</td>
<td>16.6 14.4</td>
<td>31.0</td>
<td>31.0</td>
</tr>
<tr>
<td></td>
<td>16</td>
<td>18.4</td>
<td>18.4</td>
<td></td>
</tr>
<tr>
<td>Finister</td>
<td>161</td>
<td>21.5</td>
<td>21.5</td>
<td></td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>16.9</td>
<td>16.9</td>
<td>117.8</td>
</tr>
<tr>
<td>Site</td>
<td>Sample</td>
<td>Measurement 1</td>
<td>Measurement 2</td>
<td>Measurement 3</td>
</tr>
<tr>
<td>-------------</td>
<td>--------</td>
<td>--------------</td>
<td>--------------</td>
<td>--------------</td>
</tr>
<tr>
<td>Finister</td>
<td>165</td>
<td>17.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>160</td>
<td>11.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>163</td>
<td>14.7</td>
<td>17.0</td>
<td></td>
</tr>
<tr>
<td>Maffra</td>
<td>21</td>
<td>9.8</td>
<td>6.0</td>
<td>13.9</td>
</tr>
<tr>
<td></td>
<td>31</td>
<td>13.1</td>
<td>14.4</td>
<td></td>
</tr>
<tr>
<td>Namungo</td>
<td>42</td>
<td>9.3</td>
<td>12.3</td>
<td>9.9</td>
</tr>
<tr>
<td>Numbla Vale</td>
<td>29</td>
<td>18.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>87</td>
<td>10.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>75</td>
<td>10.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tara</td>
<td>71</td>
<td>13.7</td>
<td>12.5</td>
<td></td>
</tr>
<tr>
<td></td>
<td>86</td>
<td>10.5</td>
<td>11.7</td>
<td></td>
</tr>
<tr>
<td></td>
<td>15</td>
<td>14.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tingaringy</td>
<td>53</td>
<td>12.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>62</td>
<td>14.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wullwye</td>
<td>93</td>
<td>12.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>34</td>
<td>16.3</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Determination rejected as an outlier at the 5% level.

^ Samples not considered because of demonstrable resetting.

\[
\begin{align*}
N &= 58 \\
pq &= 38 \\
p &= 13 \\
T &= 896.7 \\
T^2/N &= 13863.291 \\
\Sigma_i (T_i^2/nq) &= 14271.498 \\
\Sigma_ij (T_{ij}^2/n) &= 14419.486 \\
\Sigma_ij \alpha x_{ij}^2 &= 14508.810 \\
\end{align*}
\]

from which

\[
\begin{align*}
S_i &= 408.207 \text{ and had 12 degrees of freedom} \\
S_j(i) &= 147.988 \text{ and had 25 degrees of freedom}
\end{align*}
\]
245.

\[ S_{\alpha(ij)} = 89.324 \] and had 20 degrees of freedom.

### Table D.3

<table>
<thead>
<tr>
<th>Source of Estimate</th>
<th>Sum of Squares</th>
<th>Degrees of Freedom</th>
<th>Mean Square</th>
</tr>
</thead>
<tbody>
<tr>
<td>Between plutons</td>
<td>408.207</td>
<td>12</td>
<td>34.02</td>
</tr>
<tr>
<td>Between samples within a pluton</td>
<td>147.988</td>
<td>25</td>
<td>5.92</td>
</tr>
<tr>
<td>Between observations of one sample</td>
<td>89.324</td>
<td>20</td>
<td>4.47</td>
</tr>
</tbody>
</table>

Because \[ S_{\alpha(ij)} = \sigma^2 \]

an estimate of the standard error of one observation was

\[ \sigma = \sqrt{S_{\alpha(ij)}} = 2.11 \]

The hypothesis that there was no appreciable variation among the ages of samples from any one pluton was tested by assuming that \[ \sigma_{\lambda}^2 = 0 \]

and applying an F-test to the difference between the estimates of \( \sigma^2 \) which were calculated from \( S_{\lambda(i)} \) and \( S_{\alpha(ij)} \) respectively.

\[ F = \frac{5.92}{4.47} = 1.32 \quad F(25, 20, 0.05) = 2.08 \]

The between samples mean square was not significantly different from zero.

Once it had been established that, for the batholith as a whole, the intersample variation could be explained by variation in the observations alone, the variation in sample ages observed in any one pluton could be tested against that which was expected. The expected standard error of the mean of the observations for a pluton \((s.e._m^e)\) was given by

\[ (s.e._m^e) = \frac{2.11}{\sqrt{n_p}} \]

where \( n_p \) was the number of observations for the pluton. \((s.e._m^e)\) had \((58-1) = 57\) degrees of freedom. The observed standard error of the mean of the observations \((s.e._m^o)\) was given by

\[ (s.e._m^o) = \frac{\sigma_{\text{pop}}}{\sqrt{n_p}} \]

where \( \sigma_{\text{pop}} \) was the standard deviation of the population. \((s.e._m^o)\) had \( n_p - 1 \) degrees of freedom.
The computations for the significance tests and the results obtained are given in Table D.4.

Table D.4

<table>
<thead>
<tr>
<th>Pluton</th>
<th>(s.e.m)_e</th>
<th>(s.e.m)_o</th>
<th>F_{calc.}</th>
<th>*F_{x,y,0.05}</th>
<th>Significant?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Buckleys Lake</td>
<td>0.59</td>
<td>0.60</td>
<td>1.04</td>
<td>1.93</td>
<td>No</td>
</tr>
<tr>
<td>Cootralantra</td>
<td>1.22</td>
<td>1.48</td>
<td>1.47</td>
<td>3.17</td>
<td>No</td>
</tr>
<tr>
<td>Dalgety</td>
<td>0.75</td>
<td>0.73</td>
<td>1.06</td>
<td>3.31</td>
<td>No</td>
</tr>
<tr>
<td>Delegate</td>
<td>1.49</td>
<td>1.10</td>
<td>1.84</td>
<td>252</td>
<td>No</td>
</tr>
<tr>
<td>Finister</td>
<td>0.80</td>
<td>1.15</td>
<td>2.09</td>
<td>2.27</td>
<td>No</td>
</tr>
<tr>
<td>Maffra</td>
<td>0.94</td>
<td>1.58</td>
<td>2.80</td>
<td>2.54</td>
<td>Yes</td>
</tr>
<tr>
<td>Namungo</td>
<td>1.22</td>
<td>0.92</td>
<td>1.76</td>
<td>19.47</td>
<td>No</td>
</tr>
<tr>
<td>Numbla Vale</td>
<td>1.06</td>
<td>0.65</td>
<td>2.67</td>
<td>8.58</td>
<td>No</td>
</tr>
<tr>
<td>Tara</td>
<td>0.80</td>
<td>0.60</td>
<td>1.76</td>
<td>3.75</td>
<td>No</td>
</tr>
<tr>
<td>Wullwye</td>
<td>1.22</td>
<td>1.01</td>
<td>1.45</td>
<td>19.47</td>
<td>No</td>
</tr>
</tbody>
</table>

*x, y - degrees of freedom. Either may have been 57 or n_p - 1, depending upon whether the observed or the expected mean square was the greater.

Only for the Maffra was the range of sample ages greater than that which was expected, given the experimental error. The observation which had the largest deviation from the mean age of the Maffra was '6.0' so it was omitted and the analysis of variance repeated (Table D.5). The validity of the omission thereby was tested. Following the omission,

\[
\begin{align*}
N &= 57 \\
pq &= 38 \\
p &= 13 \\
T &= 890.7 \\
T^2/N &= 13918.359 \\
\Sigma_i(T_i^2/nq) &= 14272.490 \\
\Sigma_{ij}(T_{ij}^2/n) &= 14406.301 \\
\Sigma_{ij} x_{ij}^2 &= 14472.810
\end{align*}
\]
from which

\[ S_i = 354.131 \] and had 12 degrees of freedom.

\[ S_j(i) = 133.811 \] and had 25 degrees of freedom.

\[ S_\alpha(ij) = 66.509 \] and had 19 degrees of freedom.

Table D.5

<table>
<thead>
<tr>
<th>Source of Estimate</th>
<th>Sum of Squares</th>
<th>Degrees of Freedom</th>
<th>Mean Square</th>
</tr>
</thead>
<tbody>
<tr>
<td>Between plutons</td>
<td>354.131</td>
<td>12</td>
<td>29.51</td>
</tr>
<tr>
<td>Between samples within a pluton</td>
<td>133.811</td>
<td>25</td>
<td>5.35</td>
</tr>
<tr>
<td>Between observations of one sample</td>
<td>66.509</td>
<td>19</td>
<td>3.50</td>
</tr>
</tbody>
</table>

The estimate of the standard error of one observation was then

\[ \sigma = \sqrt{S_\alpha(ij)} = 1.87 \]

and the significance of the variation in sample ages within individual plutons had to be reassessed (Table D.6). \((s.e._m)_e\) was given by

\[ (s.e.\_m)_e = \frac{1.87}{\sqrt{n_p}} \]

and had \((57-1) = 56\) degrees of freedom.

Table D.6

<table>
<thead>
<tr>
<th>Pluton</th>
<th>((s.e._m)_e)</th>
<th>((s.e._m)_o)</th>
<th>(F_{\text{calc.}})</th>
<th>(F_{x,y,0.05})</th>
<th>Significant?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Buckleys Lake</td>
<td>0.52</td>
<td>0.60</td>
<td>1.31</td>
<td>1.93</td>
<td>No</td>
</tr>
<tr>
<td>Cootralantra</td>
<td>1.08</td>
<td>1.48</td>
<td>1.87</td>
<td>3.17</td>
<td>No</td>
</tr>
<tr>
<td>Dalgety</td>
<td>0.66</td>
<td>0.73</td>
<td>1.21</td>
<td>2.18</td>
<td>No</td>
</tr>
<tr>
<td>Delegate</td>
<td>1.32</td>
<td>1.10</td>
<td>1.45</td>
<td>252</td>
<td>No</td>
</tr>
<tr>
<td>Finister</td>
<td>0.71</td>
<td>1.15</td>
<td>2.66</td>
<td>2.27</td>
<td>Yes</td>
</tr>
<tr>
<td>Maffra</td>
<td>0.94</td>
<td>1.04</td>
<td>1.23</td>
<td>2.78</td>
<td>No</td>
</tr>
<tr>
<td>Namungo</td>
<td>1.08</td>
<td>0.92</td>
<td>1.39</td>
<td>19.47</td>
<td>No</td>
</tr>
<tr>
<td>Numbla Vale</td>
<td>0.94</td>
<td>0.65</td>
<td>2.10</td>
<td>8.58</td>
<td>No</td>
</tr>
<tr>
<td>Tara</td>
<td>0.71</td>
<td>0.60</td>
<td>1.38</td>
<td>3.74</td>
<td>No</td>
</tr>
<tr>
<td>Wullwye</td>
<td>1.08</td>
<td>1.01</td>
<td>1.14</td>
<td>19.47</td>
<td>No</td>
</tr>
</tbody>
</table>
*x, y – degrees of freedom. Either may have been 56 or n_p -1, depending upon whether the observed or the expected mean square was the greater.

The difference between (s.e.)_m and (s.e.)_e for the Maffra was no longer significant so rejection of the observation '6.0' was considered to be justified. On the other hand, the variation in the measured ages of samples from the Finister became significant. The observation which had the greatest deviation from the mean ('11.7') was omitted and the analysis of variance repeated yet again. That done, it was found however, that the omission had resulted in no significant reduction in the variation of the Finister sample ages and that the variation still exceeded expectations. For that reason, the omission was considered not to be justified.

Two hypotheses were subsequently tested.

1) That, for the batholith as a whole, the variation among the ages of samples from a pluton was no greater than the variation among the observations of the age of a sample, that is, \( \sigma_\lambda^2 = 0 \). From Table D.5

\[
\frac{F = 5.35}{3.50} = 1.53
\]

which is not significant at the 5% level \((F_{25, 19, 0.05} = 2.11)\).

2) That the variation among the ages of the plutons was no greater than the variation among the ages of samples from a pluton, that is, \( \sigma_\varepsilon^2 = 0 \).

From Table D.5

\[
\frac{F = 29.51}{5.35} = 5.51
\]

which is highly significant \((F_{12, 25, 0.01} = 2.99)\).

**SUMMARY**

From the nested classification analysis of variance the following conclusions were drawn:
1) The standard error of one observation of a rubidium–strontium biotite age from the Berridale Batholith was 1.87 m.y.

2) In the Finister Granodiorite, the variation in the measured biotite ages exceeded that expected, given the above estimate of random error.

3) For the batholith as a whole, the variation in the measured biotite ages of samples from within a pluton did not exceed that which could be accounted for by random error.

4) The difference between the measured biotite ages of plutons was highly significant, given the variation in sample ages measured within plutons.
After a correction had been made to the zircon and monazite analyses for the amounts of laboratory contamination which were measured in total-processing blanks, in all but a few cases, significant amounts of $^{204}\text{Pb}$ remained unaccounted for. The lead component which this represented was termed the 'common lead'. Since the common lead could not be analysed directly, if it was to be distinguished from the radiogenic lead in the sample, it was necessary to make an assumption as to its origin and to determine its composition by inference. Three possible sources of common lead were considered.

1) **Laboratory contamination** Several zircon studies have been done in which no distinction was made between common lead and laboratory contamination (for example, Grauert *et al.*, 1973 and Pidgeon & Hopgood, 1975). In others, although the distinction was made, the component of common lead was considered to be small in comparison to the reagent blank (Silver, 1969). This was not the case for the Berridale Batholith samples. First, the amount of $^{204}\text{Pb}$ which remained after corrections for blank varied considerably but systematically from sample to sample; the samples from each pluton had a range of $^{204}\text{Pb}$ contents which was much smaller than the total range observed. Secondly, if random contamination was the explanation for the excess $^{204}\text{Pb}$, the total-processing blanks had to be a serious underestimate both of the blank level and the extent to which it fluctuated. It was concluded therefore, that an explanation of the excess $^{204}\text{Pb}$ other than contamination was required.
2) **Magma lead incorporated at the time of crystallisation**  A very likely explanation for the common lead was that it was lead from the host magma which had been incorporated into the zircon and monazite when they crystallised. This however, proved to be unsatisfactory because when zircon inheritance was a dominant factor, all the zircons in a particular rock had not necessarily crystallised at the same time nor even within the host rock in which they were presently found. The composition of common lead in that case was difficult if not impossible to determine.

3) **Common lead from the granitoid magma**  Another possible explanation was that the common lead was derived from the magma which had formed the rock in which the zircon and monazite presently were found. This composition could be measured by the analysis of the lead in uranium-poor minerals from the zircons' host rock or from associated magmatic differentiates. Minerals which had been used in earlier studies were potassium feldspar (Tilton, 1956 and Steiger & Wasserburg, 1969 are two of many examples), galena (for example, Nier, 1939; Catanzaro & Kulp, 1964) and pyrite (Gulson & Rutishauser, 1976).

For the Berridale Batholith samples it was decided that a correction for common lead using the composition of the lead in the granitoid magma was the most appropriate. The low-uranium mineral that was chosen was potassium feldspar.

Five plutons were selected as representing the full range of granitoid chemistries in the batholith, one felsic and one mafic I-type (the Maffra and the Tara), one felsic and one mafic S-type (the Numbla Vale and the Cootralantra) and the Finister. Because
there was no means of correcting for laboratory contamination of the samples beyond that which could be estimated from total-processing blanks, the sample crushing and mineral separation was done using low-contamination techniques (see, for example, Oversby, 1975).

E.1 CRUSHING AND MINERAL SEPARATION

A sample of each rock, about 2 kg in weight, was selected for processing from the large samples which had been collected for zircon work, on the criterion that the sample must be free of veining, jointing or weathering. A hand-operated wedge splitter was used to remove all the outside surfaces and the remainder of the sample was split into 2 cm cubes.

The cubes were washed before crushing to remove any surface contamination. The washing procedure was as follows:

1) 3 min wash in 1N HCl with agitation using an ultrasonic probe.
2) Repeated rinses with 4x distilled water to remove the acid and dust.
3) Two washes of 10 min each in 4x distilled water in an ultrasonic bath.

Once the cubes had been washed, they were shattered by impact into 2-5 mm chips using a hand-operated, stainless steel piston-cylinder apparatus. The chips were ground to finer than 165μ using a Spex mixer-crusher mill and a tool steel vial. All the crushing equipment was 'contaminated'
with the sample prior to use.

The powder was sized to 85-100μm using new nylon mesh mounted in Spex plastic holders. A trial separation of potassium feldspar on this grainsize proved unsatisfactory however, so the 110-165μm fraction was separated and used instead. Before the separation was done, the dust was removed from the powder by several washes in 4x distilled water in an ultrasonic bath. The separation was done primarily by density methods, using new s-tetrabromoethane diluted with 'Pronalys' alcohol and the final magnetic cleanup which was found to be necessary was done with a Chas. W. Cook and Sons magnetic separator. The sample was washed during the separation using sub-boiling distilled acetone.

A potassium feldspar concentrate which was more than 95% pure was obtained by this method, the remainder of the sample being quartz. The additional risk of contamination which would have been involved in removing the quartz completely was not thought to be justified since the quartz was not itself a source of sufficient lead to affect the analysis.

E.2 CHEMICAL PROCESSING

Feldspar samples weighing 15 to 20 mg were analysed for uranium and lead. All reagents used, except the HClO₄ were sub-boiling distilled. All chemical processing except weighing was done in filtered air certified Class 100. The sample was dissolved overnight in a 10:1 mixture of HF and HClO₄. The dissolution reagents were dried off and the salts converted to chlorides with a few drops of 10N HCl. The sample was digested in 2 ml of 10N HCl overnight and then split approximately in half. One half was spiked with mixed ²⁰⁸Pb-²³⁵U tracer and the other kept for the lead composition determination. The spiked aliquot was kept warm overnight, dried down and then taken up in a further 1 ml of 10N HCl.

Lead and uranium were extracted from the sample using columns of AG 1-X8 200-400 mesh anion exchange resin. The resin volume used was 0.5 cc. A two stage separation was required to achieve the desired purity.
a) The sample was loaded onto the column in 1 ml of 10N HCl and the eluant from the loading plus two column washes of 1 ml, 10N HCl collected for lead. Uranium was retained by the resin. The column was washed with 2 ml of (NH$_4$)$_2$SO$_4$ solution (about 0.25N) to remove iron. The (NH$_4$)$_2$SO$_4$ was then washed off with two washes of 1 ml, 10N HCl and the uranium collected in 1 ml of H$_2$O.

b) The lead was cleaned up using the same 3.1N, 6.2N HCl column technique as was used for zircon (Appendix A).

Total processing blanks which were run with each batch of five feldspar samples were of the same order as those obtained for the zircon processing, less than 2ng total lead and less than 1ng total uranium.

The analyses of potassium feldspars from granitoids of the Berridale Batholith are listed in Table E.1 and plotted in Fig. E.1. In the calculation of the final isotopic ratios, the uranium remaining after blank corrections was assumed to be an integral part of the sample and the radiogenic lead which would have been produced in 415 m.y. was subtracted from that measured. In all cases, the correction was less than 0.25% of the sample lead.

The common lead compositions covered a much wider range in the I-type granitoids than in the S-types so it was decided that it would be unwise to use a pooled common lead composition for the correction of the zircon and monazite analyses. Instead, a common lead composition was chosen for particular granitoids on the basis of which of the measured granitoids they most resembled. The common lead compositions used are listed in Table E.2.
<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{206}\text{Pb}/^{204}\text{Pb}$</th>
<th>$^{207}\text{Pb}/^{204}\text{Pb}$</th>
<th>$^{208}\text{Pb}/^{204}\text{Pb}$</th>
<th>Pb(ppm)</th>
<th>U(ppm)</th>
<th>$^{238}\text{U}/^{204}\text{Pb}$</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Maffra Adamellite</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K 5</td>
<td>18.184</td>
<td>15.575</td>
<td>38.173</td>
<td>29.00</td>
<td>0.37</td>
<td>0.815</td>
</tr>
<tr>
<td>K 10</td>
<td>18.162</td>
<td>15.557</td>
<td>38.074</td>
<td>31.04</td>
<td>0.27</td>
<td>0.538</td>
</tr>
<tr>
<td><strong>Tara Granodiorite</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K 2</td>
<td>18.227</td>
<td>15.568</td>
<td>38.099</td>
<td>36.33</td>
<td>0.20</td>
<td>0.351</td>
</tr>
<tr>
<td>K 7</td>
<td>18.271</td>
<td>15.624</td>
<td>38.273</td>
<td>35.65</td>
<td>0.15</td>
<td>0.272</td>
</tr>
<tr>
<td><strong>Finister Granodiorite</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K 4</td>
<td>18.069</td>
<td>15.575</td>
<td>38.073</td>
<td>83.96</td>
<td>0.49</td>
<td>0.364</td>
</tr>
<tr>
<td>K 9</td>
<td>18.053</td>
<td>15.583</td>
<td>38.093</td>
<td>80.97</td>
<td>0.79</td>
<td>0.613</td>
</tr>
<tr>
<td><strong>Numbla Vale Adamellite</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K 1</td>
<td>18.097</td>
<td>15.596</td>
<td>38.116</td>
<td>72.78</td>
<td>0.45</td>
<td>0.388</td>
</tr>
<tr>
<td>K 6</td>
<td>18.055</td>
<td>15.566</td>
<td>38.011</td>
<td>72.74</td>
<td>0.56</td>
<td>0.487</td>
</tr>
<tr>
<td><strong>Cootralantra Granodiorite</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K 3</td>
<td>18.045</td>
<td>15.559</td>
<td>38.017</td>
<td>86.67</td>
<td>0.64</td>
<td>0.464</td>
</tr>
<tr>
<td>K 8</td>
<td>18.059</td>
<td>15.572</td>
<td>38.005</td>
<td>84.93</td>
<td>0.79</td>
<td>0.587</td>
</tr>
</tbody>
</table>

* Pairs of samples from each pluton are replicate analyses.

* Corrected for radiogenic lead from uranium decay.
Fig. E.1  Age corrected lead analyses of K-feldspars from the Berridale Batholith. The grid was calculated using the 'linear' Model III of Cumming & Richards (1975) which assumes a linear change in U/Pb and Th/Pb with time in order to allow for the known discrepancy in young 'model ages'.
### Table E.2

Common lead compositions

<table>
<thead>
<tr>
<th>Sample</th>
<th>$^{206}\text{Pb}/^{204}\text{Pb}$</th>
<th>$^{207}\text{Pb}/^{204}\text{Pb}$</th>
<th>$^{208}\text{Pb}/^{204}\text{Pb}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>75-117</td>
<td>18.17</td>
<td>15.57</td>
<td>38.12</td>
</tr>
<tr>
<td>75-120</td>
<td>&quot;</td>
<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>75-002</td>
<td>18.25</td>
<td>15.60</td>
<td>38.19</td>
</tr>
<tr>
<td>77-038</td>
<td>&quot;</td>
<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>74-335</td>
<td>18.06</td>
<td>15.58</td>
<td>38.08</td>
</tr>
<tr>
<td>75-107</td>
<td>18.08</td>
<td>15.58</td>
<td>38.06</td>
</tr>
<tr>
<td>77-040</td>
<td>&quot;</td>
<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>75-113</td>
<td>&quot;</td>
<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>75-109</td>
<td>18.05</td>
<td>15.57</td>
<td>38.01</td>
</tr>
<tr>
<td>77-039</td>
<td>&quot;</td>
<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>77-264</td>
<td>&quot;</td>
<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>GA834-A*</td>
<td>18.53</td>
<td>15.66</td>
<td>38.53</td>
</tr>
<tr>
<td>76-368*</td>
<td>18.12</td>
<td>15.63</td>
<td>38.05</td>
</tr>
</tbody>
</table>

* Common lead composition for an age of 176 m.y. calculated by the 'linear' Model III of Cumming & Richards (1975).

* Common lead composition for an age of 415 m.y. calculated as above.

In the case of the Dalgety Sandstone, there was no means by which the age of the common lead could be estimated so the data were corrected with both 2000 m.y. and 415 m.y. old common lead (calculated using Model III of Cumming & Richards, 1975) and the results compared. The differences were minimal and of the same order as the estimated analytical error.

When the uncertainty in the common lead correction was assessed for inclusion in the calculation of analytical errors (Appendix C) the uncertainty was considered to be equivalent to the total range of
compositions observed in all the feldspars. A crude approximation to the equivalent coefficient of variation was calculated from the range (Lyon, 1970; p. 96).

If $W$ was the range of observations, $M$ the mean observation and $n$ the number of observations, the coefficient of variation ($C_v$) was approximated by

$$C_v = \frac{(W/\sqrt{n})}{M}$$

The calculated coefficients of variation were 0.4% for $^{206}\text{Pb}/^{204}\text{Pb}$ and 0.1% for $^{207}\text{Pb}/^{204}\text{Pb}$. The contribution of error from this source to the total error in the corrected lead composition was therefore very small.
APPENDIX F

SAMPLE LOCALITIES

The samples which are listed below and located in Fig. F.1 are those from which zircons were extracted and those which have not been documented elsewhere (White et al. 1977, 1978). The locality map for biotite samples from Williams et al. (1976) has been updated (Fig. F.2).

The area which was sampled is covered by four 1:100,000 scale map sheets, the Cooma (8725), the Berridale (8625), the Numbla (8624) and the Bendock (8623). Grid references are to the 100 metre Australian map grid.

F.1  ZIRCON SAMPLES

<table>
<thead>
<tr>
<th>RSES Number</th>
<th>Field Number</th>
<th>Grid Reference</th>
<th>Rock Type and Locality</th>
</tr>
</thead>
<tbody>
<tr>
<td>77-040</td>
<td>KB42</td>
<td>8625-396997</td>
<td><strong>Bog Hole Granite.</strong> Western side of Tolbar Rd., 12 km from Eucumbene.</td>
</tr>
<tr>
<td>77-264</td>
<td>KB12</td>
<td>8625-497808</td>
<td><strong>Bullenbalong Granodiorite.</strong> Eastern side of Main Road 287, at the southern end of a small roadcut.</td>
</tr>
<tr>
<td>75-109</td>
<td>BB19</td>
<td>8625-539952</td>
<td><strong>Cootralantra Granodiorite.</strong> Northern side of the Eucumbene-Middlingbank road, 1.1 km east of Glenwood.</td>
</tr>
<tr>
<td>77-039</td>
<td>-</td>
<td>8625-686874</td>
<td><strong>Cootralantra Xenolith.</strong> Northern side of Middlingbank Rd. opposite Manderley.</td>
</tr>
<tr>
<td>75-113</td>
<td>BB9</td>
<td>8625-631837</td>
<td><strong>Dalgety Granodiorite.</strong> Northern side of Main Road 585, 1.1 km north of Cootralantra Creek.</td>
</tr>
<tr>
<td>75-120</td>
<td>BB46</td>
<td>8623-715988</td>
<td><strong>Delegate Adamellite.</strong> Southern side of Wollondibby road, 1.1 km west of Delegate.</td>
</tr>
<tr>
<td>RSES Number</td>
<td>Field Number</td>
<td>Grid Reference</td>
<td>Rock Type and Locality</td>
</tr>
<tr>
<td>-------------</td>
<td>--------------</td>
<td>----------------</td>
<td>-----------------------</td>
</tr>
<tr>
<td>74-335</td>
<td>BB163</td>
<td>8624-637347</td>
<td>Finister Granodiorite. 1.5 km west of Dalgety-Jimenbuen road, 150 m west of water tank and 30 m north of track.</td>
</tr>
<tr>
<td>75-117</td>
<td>BB31</td>
<td>8624-756536</td>
<td>Maffra Adamellite. 1.8 km west of Maffra.</td>
</tr>
<tr>
<td>75-107</td>
<td>BB2</td>
<td>8624-627414</td>
<td>Numbla Vale Adamellite. Eastern side of Dalgety-Jimenbuen road opposite turnoff to Little Popong Creek.</td>
</tr>
<tr>
<td>75-002</td>
<td>BB86</td>
<td>8625-634924</td>
<td>Tara Granodiorite. Northern side of Main Road 585, in roadcut 2.0 km northeast of Middlingbank.</td>
</tr>
<tr>
<td>77-038</td>
<td>-</td>
<td>8625-634924</td>
<td>Tara Xenolith. Same locality as 75-002.</td>
</tr>
<tr>
<td>76-368</td>
<td>-</td>
<td>8624-536546</td>
<td>Dalgety Sandstone. 100 m north of Jindabyne-Beloka road, 2.2 km west of Beloka church.</td>
</tr>
<tr>
<td>GA-834A</td>
<td>-</td>
<td>8725-914796</td>
<td>Cooma Syenite. 1.1 km west of Myalla road, hilltop 800 m west of Mandalong.</td>
</tr>
</tbody>
</table>

F.2 **RUBIDIUM-STRONTIUM TOTAL-ROCK SAMPLES**

<table>
<thead>
<tr>
<th>RSES Number</th>
<th>Grid Reference</th>
<th>Rock Type and Locality</th>
</tr>
</thead>
<tbody>
<tr>
<td>76-101</td>
<td>8623-691015</td>
<td>Delegate Aplite. Southern side of Wollondibby road, 4.5 km northwest of Delegate.</td>
</tr>
<tr>
<td>76-102</td>
<td>8623-682009</td>
<td>Delegate Aplite. 1.0 km south of Wollondibby road, western flank of hilltop 5.0 km northwest of Delegate.</td>
</tr>
</tbody>
</table>
Fig. F.1 A sketch map of the geology of the Berridale Batholith and eastern side of the Kosciusko Batholith showing the plutons and localities sampled for zircon. Aplite rubidium-strontium samples 76-101 and 76-102 also are shown. The classification of the granitoids as S- or I-type is that of White et al. (1977, 1978).
FIGURE F.1
Fig. F.2 A sketch map of the geology of the Berridale Batholith showing the plutons and localities sampled for biotite age determinations. The sample numbers are field numbers in the BB series. The classification of the granitoids as S- or I-type is that of White et al. (1977, 1978). The figure is modified from that of Williams et al. (1976).
FIGURE F.2

S-TYPE

- Cootralantra Granodiorite
- Numbla Vale Adamellite
- Tingaringy Granodiorite

I-TYPE

- Namungo Adamellite
- Tara Granodiorite
- Wullwye Granodiorite
- Maffra Adamellite
- Buckleys Lake Adamellite
- Finister Granodiorite
- Currowong Granodiorite
- Bimbimbie Granodiorite
- Delegate Adamellite
APPENDIX G

ZIRCON FRACTIONS

In the following documentation of the zircon fractions, the operating conditions of the Cook 'isodynamic' magnetic separator are expressed in terms of 'Tilt' (the tilt used so that gravity opposed the separating effect of the magnetic field), 'Slope' (the slope used to move the zircon through the magnetic field) and 'Amps' (the magnet current). Except where marked 'NM', the zircons separated were the magnetic fraction for the stated conditions. Fraction numbers for each sample correspond to those in the tables of analytical data.

Bullenbalong Granodiorite (77-264)
Starting sample - 57 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>45-85μ</td>
<td>1 (NM)</td>
<td>1°</td>
<td>20°</td>
<td>1.9</td>
</tr>
<tr>
<td>&quot;</td>
<td>2 (NM)</td>
<td>1.5°</td>
<td>20°</td>
<td>1.7</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>2°</td>
<td>15°</td>
<td>1.8</td>
</tr>
<tr>
<td>&quot;</td>
<td>4</td>
<td>1.5°</td>
<td>20°</td>
<td>1.7</td>
</tr>
<tr>
<td>30-45μ</td>
<td>M</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>NM</td>
<td>1°</td>
<td>35°</td>
<td>1.9</td>
</tr>
</tbody>
</table>

Bog Hole Granite (77-040)
Starting sample - 33.3 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100μ</td>
<td>1</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2 (NM)</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>60-100μ</td>
<td>1</td>
<td>6°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
</tbody>
</table>
### Bog Hole Granite (77-040) (Continued)

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>60-100μ</td>
<td>4 (NM)</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>60μ</td>
<td>1</td>
<td>6°</td>
<td>30°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>4°</td>
<td>30°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>30°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4 (NM)</td>
<td>2°</td>
<td>30°</td>
<td>2</td>
</tr>
</tbody>
</table>

### Cootralantra Granodiorite (75-109)

Starting sample - 10 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>100-150μ</td>
<td>M</td>
<td>2°</td>
<td>10°</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td>NM</td>
<td>2°</td>
<td>10°</td>
<td>1.9</td>
</tr>
<tr>
<td>75-100μ</td>
<td>1</td>
<td>8°</td>
<td>20°</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>7°</td>
<td>20°</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>20°</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>1°</td>
<td>26°</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td>5 (NM)</td>
<td>1°</td>
<td>26°</td>
<td>1.9</td>
</tr>
<tr>
<td>- 45μ</td>
<td>M</td>
<td>3°</td>
<td>30°</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>NM</td>
<td>3°</td>
<td>30°</td>
<td>1.8</td>
</tr>
</tbody>
</table>

### Cootralantra Xenolith (77-039)

Starting sample - 0.7 kg.

Sample divided into magnetic and non-magnetic fractions on 3° tilt, 20° slope and 2 Amps. During handpicking, each fraction was divided into two equal parts on the basis of grainsize.
Dalgety Granodiorite (75-113)

Starting sample - 31.1 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100µ</td>
<td>1</td>
<td>9°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>*4(NM)</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>75-100µ</td>
<td>1</td>
<td>9°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4(NM)</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>45- 75µ</td>
<td>1</td>
<td>9°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>4°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4(NM)</td>
<td>2°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>&lt; 45µ</td>
<td>1</td>
<td>9°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>4°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4(NM)</td>
<td>2°</td>
<td>35°</td>
<td>2</td>
</tr>
</tbody>
</table>

* Fraction further subdivided by handpicking

4' Grains with no obvious cores.

4" Grains with no obvious cores removed.

Delegate Adamellite (75-120)

Starting sample - 24.7 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100µ</td>
<td>1</td>
<td>10°</td>
<td>20°</td>
<td>0.7</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10°</td>
<td>20°</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>5°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4(NM)</td>
<td>5°</td>
<td>20°</td>
<td>2</td>
</tr>
</tbody>
</table>
### Delegate Adamellite (75-120) (Continued)

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>75-100μ</td>
<td>1</td>
<td>10°</td>
<td>20°</td>
<td>1</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>10°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>5°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>4</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>5(NM)</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>45-75μ</td>
<td>1</td>
<td>10°</td>
<td>25°</td>
<td>0.7</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>10°</td>
<td>25°</td>
<td>1</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>5°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>4</td>
<td>2°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>5(NM)</td>
<td>2°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>-45μ</td>
<td>1</td>
<td>10°</td>
<td>35°</td>
<td>0.7</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>7°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>3°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>4(NM)</td>
<td>3°</td>
<td>35°</td>
<td>2</td>
</tr>
</tbody>
</table>

### Finister Granodiorite (74-335)

Starting sample - 41.4 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100μ</td>
<td>1</td>
<td>10°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>7°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>4</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>5(NM)</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>75-100μ</td>
<td>1</td>
<td>10°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>7°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>4</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>5(NM)</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
</tbody>
</table>
Finister Granodiorite (74-335) (Continued)

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>45–75μ</td>
<td>1</td>
<td>10°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>7°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>4°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>2°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>5(NM)</td>
<td>2°</td>
<td>25°, 2</td>
<td>2</td>
</tr>
<tr>
<td>-45μ</td>
<td>1</td>
<td>10°</td>
<td>30°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>7°</td>
<td>30°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>4°</td>
<td>30°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>2°</td>
<td>30°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>5(NM)</td>
<td>2°</td>
<td>30°</td>
<td>2</td>
</tr>
</tbody>
</table>

Maffra Adamellite (75-117)

Starting sample – 37.9 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+75μ</td>
<td>1</td>
<td>10°</td>
<td>20°</td>
<td>0.9</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10°</td>
<td>20°</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>10°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>3°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>5(NM)</td>
<td>3°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>45–75μ</td>
<td>1</td>
<td>10°</td>
<td>25°</td>
<td>0.9</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10°</td>
<td>25°</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>10°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>3°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>5(NM)</td>
<td>3°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>-45μ</td>
<td>1</td>
<td>10°</td>
<td>35°</td>
<td>0.9</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10°</td>
<td>35°</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>10°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>3°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>5(NM)</td>
<td>3°</td>
<td>35°</td>
<td>2</td>
</tr>
</tbody>
</table>
**Numbla Vale Adamellite (75-107)**

Starting sample 23.9 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100µ</td>
<td>1</td>
<td>10°</td>
<td>20°</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4(NM)</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>75-100µ</td>
<td>1</td>
<td>10°</td>
<td>20°</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>5(NM)</td>
<td>2°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>45- 75µ</td>
<td>1</td>
<td>10°</td>
<td>25°</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>10°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>5°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>2°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>5(NM)</td>
<td>2°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>- 45µ</td>
<td>1</td>
<td>10°</td>
<td>35°</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>5°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>2°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>4(NM)</td>
<td>2°</td>
<td>35°</td>
<td>2</td>
</tr>
</tbody>
</table>

**Tara Granodiorite (75-002)**

Starting sample - 25 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100µ</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>75-100µ</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>45- 75µ</td>
<td>6</td>
<td>3°</td>
<td>25°</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>7</td>
<td>1°</td>
<td>25°</td>
<td>1.9</td>
</tr>
<tr>
<td></td>
<td>8</td>
<td>0.5°</td>
<td>25°</td>
<td>1.9</td>
</tr>
</tbody>
</table>
### Tara Granodiorite (75-002) (Continued)

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>45-75µ</td>
<td>9(NM)</td>
<td>0.5°</td>
<td>25°</td>
<td>1.9</td>
</tr>
<tr>
<td>&quot;</td>
<td>10</td>
<td>5°</td>
<td>25°</td>
<td>1.8</td>
</tr>
</tbody>
</table>

-45µ

After the first analyses, Fractions 7, 8 and 9 were recombined and re-subdivided to give 7', 8' and 9'. The -45µ zircon population also was subdivided.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>45-75µ</td>
<td>7'</td>
<td>2°</td>
<td>25°</td>
<td>2.1</td>
</tr>
<tr>
<td>&quot;</td>
<td>8'</td>
<td>1°</td>
<td>25°</td>
<td>2.2</td>
</tr>
<tr>
<td>&quot;</td>
<td>9'(NM)</td>
<td>1°</td>
<td>25°</td>
<td>2.2</td>
</tr>
</tbody>
</table>

### Tara Xenolith (77-038)

Starting sample - 7.5 kg.

<table>
<thead>
<tr>
<th>Size range</th>
<th>Fraction</th>
<th>Tilt</th>
<th>Slope</th>
<th>Amps</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100µ</td>
<td>1</td>
<td>7°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>3(NM)</td>
<td>4°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>70-100µ</td>
<td>1</td>
<td>7°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>5°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>3°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>4(NM)</td>
<td>3°</td>
<td>20°</td>
<td>2</td>
</tr>
<tr>
<td>45-70µ</td>
<td>1</td>
<td>7°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>2</td>
<td>5°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>3</td>
<td>3°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>4(NM)</td>
<td>3°</td>
<td>25°</td>
<td>2</td>
</tr>
<tr>
<td>-45µ</td>
<td>1</td>
<td>5°</td>
<td>35°</td>
<td>2</td>
</tr>
<tr>
<td>&quot;</td>
<td>2(NM)</td>
<td>5°</td>
<td>35°</td>
<td>2</td>
</tr>
</tbody>
</table>
Dalgety Sandstone (77-368)

The sample was divided into size fractions, +100μ, 75-100μ, 60-75μ and -60μ. The fractions were subdivided by handpicking grains of different colours.
REFERENCES 271.


BROWNE, W.R., 1931: Notes on bathyliths and some of their implications. J. roy. Soc. N.S.W., 65, pp. 112-144.


CROOK, K.A.W., 1974: Kratonization of West Pacific-type geosynclines. J. Geol., 82, pp. 24-36.


HINE, R., 1971: Granite studies of the Kosciusko Batholith, Jindabyne, N.S.W. B.Sc. (Hons) Thesis, Australian National Univ. (unpubl.).


SCHIEBNER, E., 1972a: Actualistic models in tectonic mapping. 
Rept. 24th Internat. Geol. Congress, 3, pp. 405-422.

SCHIEBNER, E., 1972b: The Kanmantoo pre-cratic province in New South Wales. 

SCHIEBNER, E., 1974: Fossil fracture zones, segmentation and correlation problems in the Tasman Fold Belt system; 
in Denmead, A.K., Tweedale, C.W. & Wilson, A.F. (Eds ), 
Aust., Qld. Division, pp. 65-98.


SHIELDS, W.R., 1960: Comparison of Belgian Congo and synthetic 'normal' samples. Table 6 in Appendix A, Report No. 8, 
National Bureau of Standards Meeting of the Advisory Committee for Standard Materials and Methods of Measurement, 


International Atomic Energy Agency, Vienna.


STEIGER, R.H. & WASSERBURG, G.J., 1969: Comparative U-Th-Pb systematics in 2.7x10$^8$ yr plutons of different geologic histories. 

STEIGER, R.H., BÄR, M.T. & BÜSCH, W., 1972: The zircon age of an anatectic rock in the Central Schwarzwald. (abstract) 


TYRRELL, G.W., 1926: The Principles of Petrology Methuen, London


RUBIDIUM-STRONTIUM AGE DETERMINATIONS ON MICAS FROM A GEOLOGICALLY CONTROLLED, COMPOSITE BATHOLITH

By I. S. WILLIAMS, W. COMPSTON, B. W. CHAPPELL, and T. SHIRAHASE

(With 2 Figures and 2 Tables)

(MS received 17 February 1975; revised MS received 23 August 1975)

ABSTRACT

Thirty biotite and two muscovite ages are reported from thirteen plutons in the Berridale Batholith, a thoroughly mapped composite batholith in southeastern Australia. Middle Ordovician country rocks place an older limit on the age of the batholith.

The mica ages of samples from any one pluton agree within experimental error except in a few cases where later plutons caused metamorphic resetting. Significant age differences have been found between plutons.

The mica ages are consistent with the mapped intrusive sequence and range between 420 m.y. and 429 m.y., which indicates cooling of the intrusives to the biotite blocking temperature of about 300°C during the Late Silurian and Early Devonian. The agreement between the muscovite and biotite ages and also one age from a total rock with high Rb/Sr suggests that this cooling interval was short.

INTRODUCTION

Southeastern Australia was the site of extensive Palaeozoic granitoid plutonism. The granitoids can be grouped into batholiths which are separated one from another by belts of country rock (Fig. 1); the batholiths in turn are composites of tens of plutons, each relatively homogeneous but in detail chemically and mineralogically distinct from its associates.

The Berridale Batholith (Fig. 2), the subject of this particular study, includes about 1700 km² of granite southwest of Cooma (Fig. 1). It has been mapped in detail and extensively sampled as part of comprehensive geochemical work, some aspects of which have been summarized by Chappell & White (1974). They used chemical, mineralogical, and field criteria to distinguish two distinct granite types, S-type and I-type, considered to result from partial melting, one of sedimentary and the other of igneous material.

S-type granites are characterized by high silica, relatively low sodium, and a high ratio of aluminium to alkalis plus calcium. This is reflected by the mineralogy. The more mafic members are very rich in biotite, sometimes containing up to 25 modal percent, whilst hornblende is absent. S-type xenoliths and sometimes the granites themselves may contain aluminosilicates, garnet, and cordierite. Chappell & White (1974) suggest that these chemical features reflect characteristics of the source material and that it has been through an earlier sedimentary cycle.

I-type granites have a much wider range in composition, but the variation in the abundance of elements within a single pluton is much more closely correlated with variation in silica content than in the S-types. I-types are relatively high in sodium and have a lower ratio of aluminium to alkalis plus calcium. Quite un-
I. S. WILLIAMS, W. COMPSTON, B. W. CHAPPELL, & T. SHIRAHASE

S-TYPE GRANITES

Cootralantra Granodiorite
West Bank Adamellite

I-TYPE GRANITES

Tara Granodiorite
Wullwye Granodiorite

Fig. 2. Plutons and localities sampled in the Berridale Batholith. The granites have been distinguished as S- or I-Type on field, petrological, and geochemical grounds. (West Bank Adamellite was renamed Namungo Adamellite while this paper was in press—Ed.)
like the S-types, the more mafic I-types commonly contain hornblende.

S-type granites are generally early in the intrusive sequence. The absolute age difference between the two types—up to a few million years or of the order of tens of millions of years—is of fundamental importance in constructing genetic models for the formation of both types of granites in single intrusive complexes.

As part of the Berridale Batholith study, a large number of rubidium-strontium whole rock analyses have been made; they will be reported elsewhere. The interpretation of the total-rock results requires that an independent estimate of the ages of intrusion be available. Consequently, in the hope that the cooling ages recorded by biotite and muscovite might correspond closely to the intrusion ages, a project to determine mica ages for the more important bodies was undertaken.

**THE PRECISION OF MICA AGES**

The plutons sampled were those for which field relationships with adjacent bodies have been reasonably well established. Freshness of sample was the prime selection criterion, and material was wherever possible obtained from several widely spaced localities within any one pluton so as to be able to recognize any internal age difference such as may result from differential cooling or later local recrystallization of the micas.

Because micas are so highly enriched in rubidium relative to strontium they may be used alone for age determinations, the result being relatively insensitive to the value estimated for the initial $^{87}\text{Sr}/^{86}\text{Sr}$. The results of analyses of micas from the Berridale Batholith are presented in Table I. The mica ages have been calculated using estimates of initial $^{87}\text{Sr}/^{86}\text{Sr}$ from total rock data. Ten duplicate analyses show that for a single measurement, the coefficient of variation for the age is $0.31\%$ (1.33 m.y.), which takes account of all random error as the great majority of the micas were analysed using the same spike. This is similar to a coefficient of variation for the age of 0.37% obtained in this laboratory for a further 14 duplicated analyses of biotites from other granites in eastern Australia (Shirahase, pers. comm.).

The measured ages are subject to systematic error due to uncertainty in the decay constant of $^{87}\text{Rb}$, here taken to be $1.39 \times 10^{-11}\text{yr}^{-1}$, and possible bias in the analytical methods. The latter is discussed in the Appendix.

**GEOLOGICAL SIGNIFICANCE**

An important consideration in the interpretation of mica ages is the ease with which biotite loses radiogenic strontium. In work on the Belt Series, northern Idaho, Hofmann (1972) demonstrated that biotite ages have been completely reset in metamorphic rocks higher in grade than the garnet zone and that some disturbance has taken place even in the biotite zone. This indicates that biotites are affected by temperatures at least as low as 450°C. Jäger & Zwart (1968) have suggested that even 300°C may be sufficient to cause biotite to become an open system for rubidium and strontium.

Whether the Berridale Batholith has been affected by regional metamorphism of sufficiently high grade to influence the biotite ages can be examined first by considering the mica ages themselves, as later metamorphism is not evident in the granites or country rocks. For the whole batholith the standard error of the mica ages is 5.1 m.y. Small though this is, the interpluton variance is significantly greater than either the variance between samples from one pluton or between duplicate analyses of the same sample. For example, analyses of seven samples from the largest pluton, the Buckleys Lake Adamellite, gave a mean age of 426.0 m.y., with 0.54 m.y. the standard error of the mean (Table I). In addition to significant differences in age between plutons, the intrusive sequence implied by the age determinations nowhere conflicts with the mapped intrusive sequence as far as it is known. These two facts are a strong argument against regional metamorphic resetting of the micas.

The Wullwye Granodiorite and Maffra Adamellite are intrusive into the Buckleys Lake Adamellite. The mean biotite age for Wullwye is 423.5 ± 1.0 m.y., which is significantly less than that for Buckleys Lake. The results for Maffra are not quite so straightforward because they are not in agreement within experimental error. This is because one biotite (sample 21) gives us an age significantly less than a second biotite and less than muscovite from the same sample. It is well known that biotite exchanges rubidium and strontium at elevated temperatures more readily than muscovite (e.g. Hanson & Gast, 1967; Baadsgaard & van Breenen, 1970), so that the anomaly within the Maffra pluton is best ascribed to a local effect on the biotite rather than to a true age difference. With the anomalous result discarded, the Maffra mica age is 422.6 ± 0.8 m.y., significantly younger than Buckleys Lake.

Table 1

<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>Age (m.y.)</th>
<th>Standard error (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>BIMBIMBIE GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>74b</td>
<td>532.4</td>
<td>16.11</td>
<td>101.0</td>
<td>1.3018</td>
<td>423.8</td>
<td>1.33</td>
</tr>
<tr>
<td><strong>BUCKLEYS LAKE ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22b(1)</td>
<td>744.1</td>
<td>13.83</td>
<td>170.7</td>
<td>1.7197</td>
<td>425.8</td>
<td></td>
</tr>
<tr>
<td>22b(2)</td>
<td>709.2</td>
<td>13.75</td>
<td>162.9</td>
<td>1.6740</td>
<td>426.1</td>
<td></td>
</tr>
<tr>
<td>33b</td>
<td>817.5</td>
<td>13.46</td>
<td>195.4</td>
<td>1.8723</td>
<td>428.0</td>
<td></td>
</tr>
<tr>
<td>10b A</td>
<td>567.3</td>
<td>7.22</td>
<td>261.3</td>
<td>2.2658</td>
<td>428.1</td>
<td></td>
</tr>
<tr>
<td>10b B</td>
<td>568.4</td>
<td>6.26</td>
<td>309.5</td>
<td>2.5511</td>
<td>427.5</td>
<td></td>
</tr>
<tr>
<td>30b</td>
<td>854.2</td>
<td>9.31</td>
<td>312.7</td>
<td>2.5613</td>
<td>425.5</td>
<td></td>
</tr>
<tr>
<td>7b A</td>
<td>786.7</td>
<td>6.78</td>
<td>417.0</td>
<td>3.2169</td>
<td>(431.8)</td>
<td></td>
</tr>
<tr>
<td>7b B</td>
<td>784.2</td>
<td>7.23</td>
<td>382.5</td>
<td>2.9756</td>
<td>425.5</td>
<td></td>
</tr>
<tr>
<td>110b A</td>
<td>843.8</td>
<td>7.73</td>
<td>384.1</td>
<td>2.9452</td>
<td>(418.1)</td>
<td></td>
</tr>
<tr>
<td>110b B</td>
<td>825.3</td>
<td>7.34</td>
<td>400.1</td>
<td>3.0801</td>
<td>425.6</td>
<td></td>
</tr>
<tr>
<td>109b</td>
<td>825.0</td>
<td>4.93</td>
<td>670.4</td>
<td>4.6677</td>
<td>423.8</td>
<td></td>
</tr>
<tr>
<td>36b A</td>
<td>543.0</td>
<td>16.40</td>
<td>101.2</td>
<td>1.3038</td>
<td>(419.6)</td>
<td></td>
</tr>
<tr>
<td>36b B</td>
<td>539.4</td>
<td>14.87</td>
<td>111.6</td>
<td>1.3688</td>
<td>(422.2)</td>
<td></td>
</tr>
<tr>
<td>83b</td>
<td>561.4</td>
<td>7.74</td>
<td>238.7</td>
<td>2.1328</td>
<td>427.0</td>
<td></td>
</tr>
<tr>
<td>3b A</td>
<td>584.7</td>
<td>9.25</td>
<td>204.0</td>
<td>1.9069</td>
<td>(420.2)</td>
<td></td>
</tr>
<tr>
<td>3b B</td>
<td>587.3</td>
<td>7.90</td>
<td>244.7</td>
<td>2.1441</td>
<td>(419.8)</td>
<td></td>
</tr>
<tr>
<td>19b A</td>
<td>641.5</td>
<td>4.82</td>
<td>496.8</td>
<td>3.7038</td>
<td>432.0</td>
<td></td>
</tr>
<tr>
<td>19b B</td>
<td>645.8</td>
<td>4.29</td>
<td>582.1</td>
<td>4.1868</td>
<td>428.2</td>
<td></td>
</tr>
<tr>
<td><strong>Cootralantra Granodiorite</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>36b A</td>
<td>543.0</td>
<td>16.40</td>
<td>101.2</td>
<td>1.3038</td>
<td>(419.6)</td>
<td></td>
</tr>
<tr>
<td>36b B</td>
<td>539.4</td>
<td>14.87</td>
<td>111.6</td>
<td>1.3688</td>
<td>(422.2)</td>
<td></td>
</tr>
<tr>
<td>83b</td>
<td>561.4</td>
<td>7.74</td>
<td>238.7</td>
<td>2.1328</td>
<td>427.0</td>
<td></td>
</tr>
<tr>
<td>3b A</td>
<td>584.7</td>
<td>9.25</td>
<td>204.0</td>
<td>1.9069</td>
<td>(420.2)</td>
<td></td>
</tr>
<tr>
<td>3b B</td>
<td>587.3</td>
<td>7.90</td>
<td>244.7</td>
<td>2.1441</td>
<td>(419.8)</td>
<td></td>
</tr>
<tr>
<td>19b A</td>
<td>641.5</td>
<td>4.82</td>
<td>496.8</td>
<td>3.7038</td>
<td>432.0</td>
<td></td>
</tr>
<tr>
<td>19b B</td>
<td>645.8</td>
<td>4.29</td>
<td>582.1</td>
<td>4.1868</td>
<td>428.2</td>
<td></td>
</tr>
<tr>
<td><strong>CURROWONG GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>48b</td>
<td>534.3</td>
<td>8.85</td>
<td>193.9</td>
<td>1.8511</td>
<td>424.3</td>
<td>1.33</td>
</tr>
<tr>
<td><strong>DALGETY GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>90b A</td>
<td>592.2</td>
<td>15.64</td>
<td>116.8</td>
<td>1.4071</td>
<td>428.6</td>
<td></td>
</tr>
<tr>
<td>*90b B</td>
<td>584.0</td>
<td>16.80</td>
<td>106.6</td>
<td>1.3445</td>
<td>427.5</td>
<td></td>
</tr>
<tr>
<td>94b</td>
<td>646.3</td>
<td>7.90</td>
<td>273.4</td>
<td>2.3176</td>
<td>422.0</td>
<td></td>
</tr>
<tr>
<td>68b</td>
<td>755.4</td>
<td>9.19</td>
<td>275.4</td>
<td>2.3425</td>
<td>425.4</td>
<td></td>
</tr>
<tr>
<td><strong>DELEGATE ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>*46b A</td>
<td>1603.4</td>
<td>26.55</td>
<td>194.0</td>
<td>1.8566</td>
<td>425.6</td>
<td></td>
</tr>
<tr>
<td>*46b B</td>
<td>796.9</td>
<td>12.04</td>
<td>214.7</td>
<td>1.9725</td>
<td>423.3</td>
<td></td>
</tr>
<tr>
<td><strong>FINISTER GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>16b</td>
<td>448.5</td>
<td>22.29</td>
<td>60.13</td>
<td>1.0659</td>
<td>427.4</td>
<td></td>
</tr>
<tr>
<td>100b</td>
<td>584.1</td>
<td>7.25</td>
<td>269.0</td>
<td>2.3047</td>
<td>425.9</td>
<td></td>
</tr>
</tbody>
</table>

In addition, the agreement of the muscovite and biotite ages implies a short time interval only between cooling to the blocking temperatures for age retention of these two minerals. The Maffra pluton may therefore have a short cooling history between emplacement and the mica age.

A second case in which the biotites do not agree within experimental error is the Cootralantra Granodiorite; two of the samples (numbers 3 and 36) are significantly younger than the rest. Cootralantra, as mapped, is a lithological unit composed of many plutons, so some dispersion in ages is not unexpected, but the
**Rb/Sr AGES FROM A GEOLOGICALLY CONTROLLED BATHOLITH**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>87Rb/86Sr</th>
<th>87Sr/86Sr</th>
<th>Age (m.y.)</th>
<th>Standard error (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>MAFFRA ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>21b A</td>
<td>1024.5</td>
<td>8.63</td>
<td>425.7</td>
<td>3.1890</td>
<td>(418.6)</td>
<td></td>
</tr>
<tr>
<td>21b B</td>
<td>1077.8</td>
<td>9.38</td>
<td>408.2</td>
<td>3.0656</td>
<td>(414.8)</td>
<td></td>
</tr>
<tr>
<td>21m</td>
<td>931.8</td>
<td>53.96</td>
<td>51.32</td>
<td>1.0073</td>
<td>422.5</td>
<td></td>
</tr>
<tr>
<td>31b (1)</td>
<td>941.8</td>
<td>10.76</td>
<td>295.8</td>
<td>2.4454</td>
<td>422.0</td>
<td></td>
</tr>
<tr>
<td>31b (2)</td>
<td>933.7</td>
<td>12.13</td>
<td>254.7</td>
<td>2.2081</td>
<td>423.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>422.6</td>
<td>0.77</td>
</tr>
<tr>
<td><strong>NUMBLA VALE ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2b A</td>
<td>891.0</td>
<td>14.26</td>
<td>201.6</td>
<td>1.9064</td>
<td>425.0</td>
<td></td>
</tr>
<tr>
<td>2b B</td>
<td>894.6</td>
<td>16.84</td>
<td>168.4</td>
<td>1.7134</td>
<td>426.5</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>425.8</td>
<td>0.94</td>
</tr>
<tr>
<td><strong>TARA GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>87b</td>
<td>481.4</td>
<td>11.78</td>
<td>126.5</td>
<td>1.4451</td>
<td>419.1</td>
<td></td>
</tr>
<tr>
<td>71b (1)</td>
<td>552.5</td>
<td>11.30</td>
<td>153.6</td>
<td>1.6110</td>
<td>422.6</td>
<td></td>
</tr>
<tr>
<td>71b (2)</td>
<td>603.7</td>
<td>9.15</td>
<td>213.8</td>
<td>1.9622</td>
<td>421.4</td>
<td></td>
</tr>
<tr>
<td>86b A</td>
<td>486.8</td>
<td>8.34</td>
<td>186.5</td>
<td>1.7965</td>
<td>419.4</td>
<td></td>
</tr>
<tr>
<td>86b B</td>
<td>487.0</td>
<td>9.49</td>
<td>161.9</td>
<td>1.6555</td>
<td>420.6</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>420.4</td>
<td>0.83</td>
</tr>
<tr>
<td><strong>TINGARINGY GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>53b</td>
<td>566.3</td>
<td>7.43</td>
<td>251.8</td>
<td>2.1884</td>
<td>420.9</td>
<td>1.33</td>
</tr>
<tr>
<td><strong>NAMUNGO ADAMELLITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>42TR A</td>
<td>200.6</td>
<td>33.16</td>
<td>17.64</td>
<td>0.80803</td>
<td></td>
<td></td>
</tr>
<tr>
<td>42TR B</td>
<td>198.1</td>
<td>33.10</td>
<td>17.45</td>
<td>0.80808</td>
<td></td>
<td></td>
</tr>
<tr>
<td>42b (1)A</td>
<td>1635</td>
<td>4.70</td>
<td>2347</td>
<td>14.384</td>
<td>418.1</td>
<td></td>
</tr>
<tr>
<td>42b (1)B</td>
<td>1633</td>
<td>4.58</td>
<td>2472</td>
<td>15.220</td>
<td>421.2</td>
<td></td>
</tr>
<tr>
<td>42b (2)</td>
<td>1621</td>
<td>5.18</td>
<td>1866</td>
<td>11.596</td>
<td>418.7</td>
<td></td>
</tr>
<tr>
<td>42m</td>
<td>863.4</td>
<td>3.48</td>
<td>1215</td>
<td>7.8340</td>
<td>420.9</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>420.1</td>
<td>0.68</td>
</tr>
<tr>
<td><strong>WULLWYE GRANODIORITE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>62b</td>
<td>801.5</td>
<td>15.84</td>
<td>159.6</td>
<td>1.6473</td>
<td>423.5</td>
<td></td>
</tr>
<tr>
<td>93b</td>
<td>1006.1</td>
<td>10.78</td>
<td>318.8</td>
<td>2.5791</td>
<td>421.7</td>
<td></td>
</tr>
<tr>
<td>34b</td>
<td>928.8</td>
<td>6.42</td>
<td>551.8</td>
<td>3.9770</td>
<td>425.3</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>423.5</td>
<td>1.04</td>
</tr>
</tbody>
</table>

b—biotite
m—muscovite
TR—Total Rock
Biotite A } Duplicate analyses
Biotite B { Duplicate analyses with (2) containing chlorite
Biotite (1) } as an impurity

Discrepancy is better explained by thermal resetting. Samples 3 and 36 were taken close to the margins of the Tara Granodiorite and Wullwye Granodiorite respectively (Fig. 2). Both these plutons intrude Cootralantra and are younger on the basis of field relations, in accord with ages reported here, Cootralantra at 428.6 ± 1.5 m.y., Tara at 420.4 ± 0.8 m.y., and Wullwye at 423.5 ± 1.0 m.y. The age for sample 3 is indistinguishable from that of Tara and sample 36 dates appreciably younger than the other samples from Cootralantra. The two samples almost certainly reflect the effects of contact metamorphism.

Because Cootralantra is composite it is unrealistic to ascribe a single intrusion age to it. However, the biotite age for at least some of the unit must be as old as 429 m.y. Further sampling would be necessary to confirm this age as representative of the entire Cootralantra Granodiorite.

One pluton in which reset biotites could not be resolved from unaffected biotites is the Finister Granodiorite. On geological grounds Finister is the oldest pluton in the batholith, being intruded by two of the large early plutons, the Numbla Vale Adamellite and Buckleys Lake Adamellite. In spite of this relationship, the biotite age for Finister (426.7 ± 0.9 m.y.) is indistinguishable from the ages of the two adjacent intrusives; 426.0 ± 0.5 m.y. and 425.8 ± 0.9 m.y. for Buckleys Lake and Numbla Vale respectively. We have not been able to resolve the problem by the use of total rock analyses and are now attempting to clarify the situation using U-Pb from zircon.

A similar inability to resolve the biotite ages of two adjacent plutons exists in the case of the Dalgety Granodiorite and Numbla Vale Adamellite, but it is quite likely that this simply represents the limit of resolution of the method. The biotite age of 425.2 ± 1.8 m.y. for Dalgety is indistinguishable from 425.8 ± 0.9 m.y. for Numbla Vale, even though Dalgety plainly intrudes Numbla Vale.

The Namungo Adamellite is intrusive into the Cootralantra Granodiorite, but its relationship to the Tara Granodiorite is not yet resolved. Namungo is a small pluton containing garnet, andalusite, and muscovite; it has a particularly high ratio of rubidium to strontium. Its mica age, 420.1 ± 0.7 m.y., is indistinguishable from that of Tara but appreciably younger than Cootralantra. Because of its high present day ⁸⁷Sr/⁸⁶Sr, the age of the total rock sample for Namungo (Table I) is rather insensitive to the initial ⁸⁷Sr/⁸⁶Sr, for which 0.705 may be taken as the lowest reasonable value. The result is a total rock age of 421 ± 1.3 m.y. (s.e.), which represents a maximum possible value if the total-rock has remained a closed chemical system. Thus the Namungo pluton is unique in the Berridale Batholith in having concordant and sensitive biotite, muscovite, and total-rock ages. For this pluton the time difference between emplacement, as registered by the total-rock age, and the cooling of biotite cannot exceed about 3 m.y. This conclusion can be avoided only by making the ad hoc assumption that the Namungo total-rock sample and its micas were completely reset in their Rb-Sr ages during the emplacement and cooling of the Tara Granodiorite.

**STRATIGRAPHICAL CONSTRAINTS TO THE AGES**

The internal consistency demonstrated above is not a sufficient test of the accuracy of the age determinations; they must also be consistent within any age constraints placed on intrusion by fossils in the country rocks. Graptolites are abundant at several localities near the batholith (Sherrard, 1953) and, except for one Gisbornian assemblage, are all of Eastonian age. This does not preclude the possibility that some of the sedimentary sequence is older.

The Early Palaeozoic time scale is only poorly linked with geochronology, a problem towards which some current work in this laboratory is directed. Estimates of the age of the Caradoc (the approximate British equivalent of the Australian Eastonian) have been made by several authors but are in great need of refinement.

Byström *et al.* (1961) worked on bentonites from Kinnekulle, Sweden, which are firmly located in the early Caradocian. K-Ar dating of both biotite and sanidine gave a mean age of 453 ± 20 m.y. (recalculated from 444 m.y. using λβ = 4.905 × 10⁻¹⁰ yr⁻¹; λ, + λ/β = 5.747 × 10⁻¹¹ yr⁻¹ [Beckinsale & Gale, 1969]). This age is supported by determinations on the early Caradocian Bail Hill Volcanics (Harris *et al*., 1965). K-Ar analyses of biotites from this mica andesites carried out in Toronto, Oxford, and Cambridge yield a mean age (recalculated from 445 m.y. using the Beckinsale & Gale constants) of 456 ± 11 (2σ) m.y. Harris also reports a K-Ar mean of 485 ± 6 (2σ) m.y. (recalculated from 475 m.y.) on gabbro which he considers to be intrusive into, but broadly contemporaneous with, middle Arenigian lavas at Ballantrae, thus implying this to be a minimum age for that period. In fact, the middle Arenigian could be considerably older.

Bofinger *et al.* (1968) report 475 ± 3 m.y. for the age of burial metamorphism in Middle to Upper Ordovician shales, based on internal Rb-Sr isochrons generated by selective leaching of secondary minerals. The age employs the same decay constant as the present paper (λ ⁸⁷Rb = 1.39 × 10⁻¹¹ yr⁻¹), and is referenced to an age of 1424 ± 5 m.y. for the standard K-feldspar NBS 70a. As the reference sample used with the Berridale micas, NBS 607, is a sized fraction of the earlier NBS 70a, the 475 m.y. of Bofinger *et al.* should be
Rb/Sr AGES FROM A GEOLOGICALLY CONTROLLED BATHOLITH

adjusted to 483 m.y. according to the fractional difference between 1424 m.y. and the 1447 m.y. reported here for NBS 607. Thus the Eastonian probably extends to at least 480 m.y. A similar conclusion was reached by Poole et al. (1963), who proposed 488 m.y. as a younger limit for the Caradocian, although this has been disputed on geological grounds by Rickard (1964), who interpreted the age as an older limit.

The biotite ages from the Berridale Batholith, lying between 420 and 430 m.y., do not conflict with the age of the country rocks. That is not to say, however, that the biotites are recording the actual intrusion age, because only in the Namungo Adamellite can we yet distinguish between the times of emplacement and of cooling to below the $\sim$300°C blocking temperature for the biotite ages. It is possible, in fact, that the older plutons within the Batholith may be as old as the stratigraphical limit, say 475 m.y., and either that their biotites have been reset by the younger plutons, or that cooling of the older plutons to 300°C extended from the (hypothetical) 475 m.y. to about 425 m.y.

Bofinger et al. (1970) have reported a Rb-Sr biotite age of 438 ± 4 m.y. for Upper Silurian volcanics ($\lambda^{87}$Rb = 1.39 × 10^{-11} yr^{-1}), which serves to relate the ages of the Berridale micas to the geological time-scale. The identical biotites used by Bofinger et al. were reanalyzed by Roddick (1974), using the same calibrated spikes as the present work, to confirm a suspected analytical bias (de Laeter et al. 1973) in the 438 m.y. result.* Roddick obtained 430 ± 4 m.y., in good agreement with the expected bias, and we take this revised value for the age as the best available estimate for the Late Silurian (Ludlovian). It follows that the oldest of the Berridale micas are Late Silurian, so that the oldest, and some of the intermediate-age, plutons are Silurian or older and had cooled by the Late Silurian. However, the youngest plutons were probably both intruded and cooled in the Devonian.

CONCLUSIONS
1. The biotite ages for 13 plutons examined from the Berridale Batholith are set out below.
2. Our present data suggest a relatively short time interval, less than 10 m.y., for the emplacement and crystallization of the spectrum of rocks in the Berridale Batholith during the Late Silurian and Early Devonian. It is possible that the older S-type plutons such as the Cootralantra Granodiorite have had their mica ages reset by the younger I-type plutons. It appears that such resetting can be observed in the vicinity of younger plutons; ages

<table>
<thead>
<tr>
<th>Biotite age ± 1σ (m.y.)</th>
<th>Field Relationship</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>Oldest</td>
</tr>
<tr>
<td>Finister Granodiorite</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td></td>
</tr>
</tbody>
</table>

* Bofinger et al. (1970) state that their age results are referenced to spike calibrations that give 530 ppm Rb and 3.00 ppm radiogenic Sr for NBS 70a K-feldspar, which is equivalent to an age measurement of 1427 m.y. This means that their 438 m.y. result should be low by about 2.0% compared with present data, rather than high as observed. Re-examination of the relevant calibrations shows that Bofinger et al.'s (1970) statement refers only to the data in their Table 3 (shales) and not their Table 1 (igneous rocks and biotites), for which a different spike that was not calibrated with respect to the K-feldspar standard was used. Bias in the Bofinger et al. (1970) biotite ages should be judged therefore solely on the results cited here for the repeat analyses of their biotites.

The results of Bofinger et al. (1968) are correctly referenced to simultaneous direct measurements of NBS 70a that yielded 1424 m.y. (recalculated to 1.39 × 10^{-11} yr^{-1} for $\lambda^{87}$Rb).

measured on samples further removed from younger plutons are significantly older and may closely approximate the age of crystallization. It is possible nevertheless that the oldest plutons are substantially older than their biotite ages, these having been uniformly reset by the intermediate age plutons. If so, their maximum age would be Middle Ordovician, according to the fossiliferous metasediments that they intrude.

4. Muscovite ages are in exact agreement with biotite ages for the Maffra and Namungo Adamellites, implying a short cooling history for these bodies. Their mica ages are probably close to their emplacement ages.

5. The Namungo Adamellite alone, of all the plutons, has a total rock age that is not sensitive to the value for initial 87Sr/86Sr. It agrees well with the mica age. Unless the total rock has been an open chemical system (for which we have no evidence), the mica age is, within experimental error, equal to the emplacement age for this pluton.

APPENDIX

Analytical details
About 80 mg of sample was prepared for analysis by dissolution in hydrofluoric and perchloric acids, conversion of the salts to chlorides with hydrochloric acid, and spiking with mixed 84Sr/88Sr tracer. Rubidium and strontium were concentrated using cation exchange resin. The use of weighed aliquots is not necessary. The use of weighed aliquots is not necessary and the errors associated with doing so are avoided. Errors in weighing both the sample and tracer cancel out for the measurement of 87Rb/86Sr, and evaporation does not change 86Rb/88Sr in the tracer.

The latter was calibrated using the stoichiometric Rb and Sr salts SRM 984 and SRM 987 of the U.S. National Bureau of Standards.

Replicate measurements of the standard K-feldspar NBS 607 (Table II) using the same spike and mass spectrometer gave a mean 87Rb/86Sr of 24.18 ± 0.04 (s.e.m) and a mean 87Sr/86Sr of 1.20114 ± 0.00015, equivalent to an age of 1447 m.y. using 0.7100 as the initial 87Sr/86Sr. This value for age serves as an indicator for systematic error in the overall measurement operation. It is within 0.35% of the age reported by the U.S. National Bureau of Standards (1442 m.y.), a difference of the same order as our random error (0.3%).

<table>
<thead>
<tr>
<th>Table II</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Replicate Rb, Sr analyses of SRM 607, the reference potassium feldspar</strong></td>
</tr>
<tr>
<td>Rb (ppm)</td>
</tr>
<tr>
<td>---------</td>
</tr>
<tr>
<td>1A</td>
</tr>
<tr>
<td>1B</td>
</tr>
<tr>
<td>1C</td>
</tr>
<tr>
<td>2A</td>
</tr>
<tr>
<td>2B</td>
</tr>
<tr>
<td>2C</td>
</tr>
</tbody>
</table>

24.18 ± .004 (s.e.m) 1.20114 ± 15 (s.e.m)

REFERENCES


T. Shirahase, 
*Geological Survey of Japan* 
(Visiting Fellow, Research School of Earth Sciences, Australian National University).

---

J. S. Williams, 
W. Compston, 
*Research School of Earth Sciences,* 
B. W. Chappell, 
*Department of Geology,* 
*Australian National University.*