STRUCTURE AND TECTONO-THERMAL HISTORY OF THE EASTERN MOUNT Isa INLIER, AUSTRALIA

by

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STATEMENT

This thesis is based on fieldwork and numerical modelling, carried out respectively in the Mount Isa Inlier, Queensland, and the Department of Geology, Australian National University, between June 1985 and July 1988. An introductory field season in 1984 preceded the Ph.D. project.

In the introductory field season, Mr. Guido Schreurs from the State University of Utrecht, the Netherlands, accompanied me to the Deighton and West-Leichhardt areas. In 1985, I revisited these areas, reinterpreted the field data, and wrote a paper for the "Australian Journal of Earth Sciences". A reprint of this paper is presented in Appendix 1. Schreurs, having participated in the introductory fieldwork, was a junior author.

In 1985, image-processing of confidential aero-magnetic and aero-radiometric datasets was carried out in cooperation with Mr. Wayne Stasinowski from the Broken Hill Proprietary, Co. Ltd., Camberwell.

In the final stage of the project, discussions with Dr. Mike A. Etheridge about a wider application of the numerical results led to the submission of a short paper to "Nature", with Etheridge as junior author (Appendix 8).

Unless otherwise acknowledged, all analyses, arguments and conclusions presented in this thesis are my own.

Ramon J.H. Loosveld
Science is built up with facts as a house is with stones.
But a collection of facts is no more science than a heap of stones is a house.

Henri Poincaré (1854-1912)

and

To specialize is to brush one tooth.

Tom Robbins
(in "Even cowgirls get the blues", 1977)


to INEKE
ACKNOWLEDGEMENTS

During the course of my studies at the ANU, I was in receipt of an ANU Ph.D. Scholarship. Additional financial assistance came from the Broken Hill Proprietary Co. Ltd., Brisbane (BHP), and the Bureau of Mineral Resources, Geology and Geophysics (BMR). The latter two also provided logistical support in the field, in the form of vehicles, maps and other equipment, and field hands. Additionally, BHP provided both raw and processed magnetic and radiometric data. To all three institutions I am grateful.

I wish to thank my supervisors, Dr. Mike J. Rickard, Dr. Mike A. Etheridge (1986-1988), Dr. Gordon S. Lister (second half 1985), and Dr. Dave J. Ellis (second half 1987; the rest of the time he was advisor to the project). All four supervisors/advisors gave me almost unlimited freedom in defining a suitable problem and tackling it. I thank them for their confidence. Discussions with them are greatly appreciated, as well as their criticism on the (many) manuscripts. I am also grateful to Mike Rickard for introducing me to the Lachlan Fold Belt.

This study builds heavily on the large database constructed by BMR workers over the last two decades. I am indebted to the following co-workers of BMR's Mount Isa Regional and Tectonic History (MIRTH) project for freely-given information and discussion: Dr. Dave Blake, Dr. Ken Erikson (also Virginia Polytechnic Institute, VA, USA), Ms. June Hill, Mr. Jim Jackson, Dr. Rod Page, Dr. Rod Ryburn, Dr. Alistair Stewart, Dr. Peter Williams, Dr. Lesley Wyborn. The study also benefitted from by discussions with: Ms. Diane Bettess (Monash University), Dr. Warrington Cameron (ANU), Mr. Ken Chapple (BHP), Dr. Rod Holcombe (Queensland University), Dr. Bill Laing (James Cook University), Dr. Steve Lonker (ANU), Mr. Nick Oliver (Monash University), Dr. Cees Passchier (Utrecht University, the Netherlands), Mr. Paul Pearson (Queensland University), Mr. Bill Perkins (Mount Isa Mines), Mr. Jürgen Reinhardt (James Cook University), Dr. Mike Rubenach (James Cook University), Mr. Bob Skrzeczyński (BHP), Mr. Wayne Stanisowski (BHP), Mr. Peter Stuart-Smith (ANU), and Dr. Steve Walters (BHP).

Three manuscripts, which are incorporated in this thesis, had been reviewed for journal publication at the time of submission. I am indebted to the reviewers Drs. Rod Holcombe and Gordon Lister (Appendix 1), Drs. N. Rast and M.P. Ryan (Chapter 4) and Dr. Alison Ord and two anonymous reviewers (Chapter 3).
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Over the last nine (!) years (my B.Sc., M.Sc., and Ph.D. periods), I had the privilege of working with Gordon Lister. He invited me to Australia in 1984 to participate in the MIRTH project, lured me into this Ph.D. project, and, until the end, remained a helpful and inspiring supervisor/advisor and a good friend.

Of the technical staff of the Department of Geology at the ANU, I wish to particularly thank Chris Foudoulis for his assistance with photography and XRD analyses, and Robin Westcott for the preparation of my thin sections. The staff of A.N.U.'s Science Photographic Unit was always patient, helpful, and efficient. Roland Aronsen helped with some of the illustrations. Miss Dianne Pillinger (BMR) drafted appendix 9.

Finally, I thank my friends (especially Jurriaan Gerretsen and Bill Kiene) for their fraternal advise and good times shared, my parents for always encouraging me to study, even when it was at the antipodes, and Ineke, not only for her assistance and companionship in the field, or for being a far more patient programmer than I ever will be, but especially for her everyday support, love and patience. Thank you, Ineke!
ABSTRACT

This thesis reports on (1) a field study of the central eastern Mount Isa Inlier, NW Queensland, Australia, and (2) a theoretical study of the synchronicity of low-pressure facies metamorphism and crustal thickening. The Mount Isa Inlier has previously been considered to represent either (a) the eastern margin of a Middle Proterozoic continent, or (b) an entirely ensialic mobile belt. In this (field) study, the tectono(-thermal) history of the central eastern part of the inlier, is examined. Three major tectono-stratigraphic domains are recognized. All three underwent the same major tectonic events, although the style of deformation is different in the three domains. The combined structural framework consists of two extensional events at about 1780 Ma. (D\textsubscript{e1}) and 1750 Ma. (D\textsubscript{e2}), and at least three compressional events, probably at 1610 Ma. (D\textsubscript{c1}), 1550 Ma. (D\textsubscript{c2}), and 1510-1450 Ma. (D\textsubscript{c3}). These three compressional and two extensional events are correlated with similar events in the central and western parts of the inlier. A third extensional event affected the western part of the inlier at about 1680 Ma. (D\textsubscript{e3}).

The basin-modifying events in the eastern part of the inlier are characterized by early thrusts (D\textsubscript{c1}), followed by penetrative, N-S trending, upright folds (D\textsubscript{c2}), and a wide range of late structures ranging from strike-slip and reverse faults to WNW-trending, upright folds (D\textsubscript{c3} or D\textsubscript{c3+}). Major D\textsubscript{c1} structures are: the Snake Creek Anticline, the Toole Creek and Oonoomurra Synclines and the Pumpkin Gully Thrust; major D\textsubscript{c2} structures are: the Mitakoodi Anticlinorium, the Weatherly Creek Syncline, the Middle Creek Anticline and the Pumpkin Gully Syncline; major D\textsubscript{c3}(+) structures are: the Cloncurry "Overthrust", the WNW-trending fold around which the Pumpkin Gully Syncline is folded, and the shear zone between the Mitakoodi Anticlinorium and the Tommy Creek Block. Mostly, D\textsubscript{c3}(+) structures cannot be correlated with each other.

Having established the structural framework, the relation between the three tectono-stratigraphic domains becomes clearer, and a new stratigraphic correlation scheme is put forward. There is insufficient litho-stratigraphic evidence to maintain the continental margin model, which was based on the interpretation of an eastward-deepening basin. The new litho-stratigraphic framework supports the alternative hypothesis of ensialic basin development (between about 1800 and 1650 Ma.), previously proposed on the grounds of the local presence of continental basement, and the absence of ophiolites, andesites, and high-pressure metamorphic assemblages.

The penetrative event (D\textsubscript{e2}), in which the crust was probably thickened (=compressed) by a factor of 1.5 to 2.2, was accompanied by prograde low-pressure facies metamorphism. This is in stark contrast to results from numerical experiments in which thickening of the entire lithosphere, or thickening of the crust with no change in
lithospheric thickness, was simulated (England & Thompson, 1984): in these, crustal thickening, under a wide range of reasonable parameterisations, will result in essentially adiabatic compression (high-P facies metamorphism), with temperatures increasing during the later stages of, and after, the crustal thickening. Early tectonic denudation and/or erosion may interfere with this temperature rise (-ΔP, +ΔT), but the later stages of decompression are invariably characterized by decompressive cooling (-ΔP, -ΔT).

Thus, the classical numerical models correlate well with the clockwise P-T-t trajectories (in the P-T field) as petrologically deduced from many Alpino-type orogens. In the Mount Isa Inlier, however, as in many other Australian Early to Middle Proterozoic fold belts, prograde low-P facies metamorphism (+ΔP,+ΔT) was succeeded by essentially isobaric cooling. Thus, its metamorphic history is typified by an anti-clockwise P-T-t path.

Low-P facies metamorphism is generally attributed to lithospheric extension and/or magmatic processes, either mafic underplating or crustal melting. Both magmatic processes can induce prograde low-P (andalusite-sillimanite) facies metamorphism, albeit of restricted temporal and spatial extent. Considering the distribution of granitoids, their emplacement ages, and the isograd pattern, it can be concluded that magmatism plays a negligible role in the development of andalusite/sillimanite metamorphism in the Mount Isa Inlier. Granitoids enriched in heat producing elements, however, substantially contribute to the local heat balance.

More importantly, both mafic underplating and crustal melting require a substantial, thermal anomaly in the upper mantle. This study is primarily concerned with the fundamental cause of this thermal anomaly, upon which the magmatic effects can be considered second order. Two such fundamental causes are envisaged and modelled here: (1) lithospheric extension, and (2) crustal thickening accompanied by convective thinning of the mantle lithosphere. Both have been numerically simulated in various modes and with a wide range of parameters, using a one-dimensional, finite-difference approach. Both can explain anomalously low-P facies metamorphism. Rapid and strong lithospheric thinning, however, cannot cause prograde (+ΔT) low-P metamorphism during a later (>=5 Ma.) phase of crustal thickening. In the Mount Isa Inlier, the time interval between the extensional events and low-P facies metamorphism significantly exceeds the thermal time constant for thermal relaxation of a thinned lithosphere: in other words, the depth-temperature profile would have readjusted by the time crustal thickening commenced. Also, as the numerical modelling shows, the amount of heat released by elevation and/or compression of isotherms during extension will only be sufficient to explain the observed metamorphic assemblages if finite extension was about 300%. Extension factors in the inlier, however, were probably lower.
The second, and favoured, model involves crustal thickening associated with the synchronous upwelling of hot asthenospheric material, as theoretically modelled by Houseman et al. (1981). If this model is to explain the documented anti-clockwise P-T-t paths, my numerical experiments place constraints on (a) the steady-state conductive geotherm, (b) the intensity of the deformation, (c) the strain rate of the thickening event, (d) the duration of convection in the newly-upwelled high-level asthenospheric material, and (e) the overlap in time between the crustal thickening event and the thinning of the mantle lithosphere. It is concluded that prograde low-P facies metamorphism can accompany the later stages of crustal thickening, if the latter is protracted (low strain rates or multiple events). As time constants for erosion are generally much greater than thermal time constants, thickening will be succeeded by slightly decompressive cooling (assuming crustal thickening and convection within the newly-upwelled asthenospheric material stop at the same time).

Numerical experiments simulating the complete removal of the mantle lithosphere during the early stages of low strain-rate crustal thickening and whose parameterisation includes the anomalous enrichment of radioactive isotopes in mid-crustal slabs, like in the pre-metamorphic granites of the inlier, yield P-T-t paths similar to those petrologically deduced for the Mount Isa Inlier and other Middle Proterozoic fold belts in Australia (especially the Willyama Province).
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CHAPTER 1

INTRODUCTION

LOOSVELD
INTRODUCTION

1.1 INTRODUCTION

The theory of plate tectonics provides an elegant explanation for the formation and deformation of oceanic crust, and is now widely-accepted. Consensus on the growth and deformation of continental crust on the other hand is still far-off. Questions definitely not fully answered are: is continental accretion pre-dominantly subduction-related, and thus restricted to destructive continental margins, or does underplating of mantle material in an intra-plate situation play an important role? Is there a secular change from the latter in the Archean to the former at present? How does the continental lithosphere deform? In this thesis a model for ensialic orogeny is put forward, based on a reconstruction of events that affected the Mount Isa Inlier, NW Queensland, Australia (Fig. 1.1).

Fig. 1.1 Distribution of Australian Archaean cratons and Early to Middle Proterozoic fold belts (from Etheridge et al., 1987b).

The Mount Isa Inlier is located between latitudes 18 and 22°S and longitudes 138°30' and 141°00'E, in the northwestern part of Queensland. Aeromagnetic and gravity data show it to be of considerable greater extent in the subsurface (Blake's 1987 figure 2). The inlier has, mainly due to the occurrence of gold and copper deposits,
Fig. 1.2 Schematic map of the Mount Isa Inlier (from Blake, 1986). The Lawn Hill Platform and the Leichhardt River Fault Trough broadly correspond to the Western Fold Belt, the Kalkadoon-Leichhardt Block to the Kalkadoon-Leichhardt Belt, and the Wonga-Duchess Belt, Malbon Block and Mary Kathleen Shelf to the Eastern Fold Belt.
attracted geological attention since the late 19th century (Rands, 1895; Jack, 1898). Major progress was made with the introduction of aerial photography in 1936 (Honman et al., 1939). The Soldiers Cap Group area, in the east of the inlier (Fig. 1.2), was in fact the first area in Queensland to be covered by aerial photography. The first comprehensive study of the inlier was carried out by Carter et al. (1961), which resulted in (preliminary) one inch to four miles geological map coverage of the complete inlier. As part of the same Bureau of Mineral Resources/Geological Survey of Queensland (BMR/GSQ) program, Carter (1959) produced the CLONCURRY 1:250 000 Sheet (2nd edition by G.S.Q., 1983) and Carter & Öpik (1963) produced the Duchess 1:250 000 Sheet. The last one and a half decades have seen a 1:100 000 mapping program covering the inlier. This program is now in its last phase with the preparation of the final 1:100 000 CLONCURRY Sheet (Ryburn et al., in prep.). The (preliminary) 1:100 000 map commentaries commonly give the most comprehensive data compilation (e.g. Derrick et al., 1971).

The Mount Isa Inlier is part of an Early to Middle Proterozoic (2200-1400 Ma.) fold belt, unconformably overlain and surrounded by the Paleozoic Georgina Basin in the south and west, the Mesozoic Carpentaria Basin in the north and north-east, and the Eromanga Basin in the south-east (Fig. 1.2). Its present crustal thickness is approximately 40 km (Wellman, 1976; Shirley, 1979; Dooley, 1980; Finlayson, 1982). The inlier forms a broad gravity high (Bouguer and Free Air), the gradual increase at its margins indicating that the anomalous mass lies at depth. This anomalous mass, which also shows up on deep refraction seismics (Drummond & Collins, 1986), probably is a mafic underplate, and can be the source of the younger, more evolved felsic products in the inlier. This underplate is reflected in a widespread Sm-Nd signature of approximately 2000+ Ma. throughout the Early to Middle Proterozoic of Australia (McCulloch, 1987).

Both the structural and litho-stratigraphic grains of the inlier are N-S. Structurally, the inlier is characterized by upright, N-S trending folds (of all scales), compartmentalised by younger NNE- and NNW-trending strike-slip faults and high-angle reverse faults. Litho-stratigraphically, Carter et al. (1961) recognized three broad belts: a western and an eastern succession, separated by the Kalkadoon-Leichhardt basement block. Blake (1987) upgraded this concept of different belts to a tectono-stratigraphic subdivision by defining the Western Fold Belt, the Kalkadoon-Leichhardt Belt and the Eastern Fold Belt. Apart from the BMR regional mapping programmes, the Western Fold Belt, which hosts the Mount Isa Cu/Pb/Zn/Ag deposit, has received markedly more attention than the central and eastern belts.
1.2 ACTUALISM IN THE PROTEROZOIC OF AUSTRALIA?

In the "Proterozoic" research community, a polarized discussion exists between those who support actualistic models of plate tectonics (e.g. Windley, 1981; Hoffman, 1973, 1980) and those who visualize a non-uniformitarian variant (e.g. Kröner, 1983; Rutland, 1976, 1982; Etheridge et al., 1987b). For the Australian continent the problem was defined as follows (Etheridge & Wybom, 1988):

"...the continent evolved by lateral accretion of terranes to Archaean nuclei, which fragmented in one or more periods of plate motion similar to modern tectonic settings [....or...] ...the various [Early to Middle Proterozoic] provinces developed largely on Archaean sialic basement in their present relative locations, without significant continental fragmentation by sea-floor spreading...".

The former viewpoint is classified as "mobilist", the latter as "fixist". The paleomagnetic evidence favours the fixist's viewpoint, as a single apparent polar wander path for Australia can be drawn through the poles, even though these poles are derived from different cratonic units (Idnurm & Giddings, in press). It is not conclusive though.

This controversy also appears in the literature on the Mount Isa Inlier. Plumb & Derrick (1975), Dunnet (1976b), Wilson (1978), and Plumb et al. (1980) have argued that the Mount Isa Inlier represents the eastern margin of a Middle Proterozoic continent. This theory was supported by four main lines of evidence. 1) Based on a limited geochemical database, Glikson et al. (1976) recognized an eastward change in mafic volcanic chemistry from continental tholeiites in the west to ocean floor tholeiites in the east. Wilson (1978) favourably compared the felsic volcanics with Andean-type rhyolites. He reported eastward decreasing K$_2$O/SiO$_2$ values. 2) Rocks eastwards of the inlier were all supposed to be younger. 3) Deepwater sediments, notably turbidites, are restricted to the eastern part of the inlier. 4) Eastward thinning of the crust under the inlier was inferred from an assumed increase in ductility and metamorphic grade to the east, and from a regional negative Bouguer anomaly to the east of the Mount Isa Inlier.

However, Wybom & Blake (1982) convincingly showed that none of these four arguments may be valid. 1) Based on a considerably larger chemical database, they concluded that mafic volcanics both in the west and east are of continental tholeiitic affinity. Subsequent work on dolerites by Ellis & Wybom (1984) and Hamilton (1985) also indicated a continental source. Also, none of the four main felsic volcanic suites (ranging in age from 1860 to 1680 Ma.) show eastward-decreasing K$_2$O contents. Andesites are either absent or very scarce in the inlier. 2) Scattered outcrop and drillhole data show that Precambrian granites and metamorphics, and no oceanic crustal sequences, are buried under the Mesozoic Eromanga Basin east of the inlier. Laing & Beardsmore (1986) now correlate the turbiditic sequence of the eastern Mount Isa Inlier with the upper Etheridge Group of the Georgetown Block (Fig. 1.1). 3) Wyborn &
Blake (1982) state that the average grain size and feldspar content of the eastern turbidites decrease westwards, indicating a source towards the east. 4) Amphibolite-grade metamorphics occur throughout the inlier, not just in the east. Ductile behaviour is related to rocktype and metamorphic grade, and not to lateral position in the inlier. Moreover, there cannot possibly be a direct link between the regional metamorphism of an age of ≈1550 Ma. (Page & Bell, 1986) and the extrusion of the volcanic suites of an age range of 1860 to 1680 (or 1620) Ma. Finally, the regional gravity maps (BMR, 1976; Blake's 1987 figure 2) show a regional Bouguer low both east and west of the inlier, and are thus not indicative of overall eastwards crustal thinning.

In the Mount Isa Inlier, the strong bimodality of igneous rocks, i.e. basalt and dolerite versus rhyolite, dacite and granite (Bultitude & Wyborn, 1982), point towards an ensialic (rifting) origin rather than to an origin related to the formation and subsequent destruction of oceanic crust (compare e.g. with Rio Grande Rift; Riecker, 1979). A review of the development of the "ensialic rifting" concept for the inlier is given by Beardsmore et al. (in press). In the last 6 years, increasing evidence has surfaced for an entirely ensialic origin, not only for the Mount Isa Inlier, but for all northern Australian Early to Middle Proterozoic fold belts (Etheridge et al., 1984b, 1987b). The only significant difference between these fold belts and modern-day plate tectonic analogues appears to be that in the former continental extension did not lead to the formation of oceanic crust. The Mount Isa Inlier has become the best documented example of an orogen with entirely ensialic characteristics. Unfortunately, this appears to be the only conclusion most workers in the inlier agree upon. Considering the diversity of proposed models for the evolution of the inlier, this is even more remarkable.

To resolve some of the differences, the BMR, in 1984, chose to further update the database of the Mount Isa Inlier, and further test the non-actualistic model, as evolved within the BMR (Etheridge et al., 1984b, 1987b): a special research project, the Mount Isa Regional and Tectonic History project (MIRTH), supervised by Drs. David Blake and Gordon Lister (now Monash University), was set up. The following geological framework of the inlier emerged from the MIRTH project. This Ph.D. study was part of the MIRTH project.

1.3 GEOLOGICAL SETTING OF THE MOUNT ISA INLIER

1.3.1 Structure

The earliest tectonic events remain enigmatic. Pre-1875 Ma. basement inliers unconformably underlie two or three cover sequences (Fig. 1.3). These cover sequences are volumetrically dominated by shelf sediments: huge amounts of fine-grained
Introduction

KALKADOON-LEICHHARDT EASTERN FOLD BELT

- Granite
- Microgranite
- Gabbro
- Felsic volcanics
- Mafic volcanics
- Lower crustal mantle derived underplates (with age of underplating)
- Recognised unconformity

DEFORMATIONS AND METAMORPHISM

1810 - 1550 Ma
(calcsilicate) sediments are interbedded with shallow marine quartzites (e.g. Mary
Kathleen Group) and shales (Mount Isa and McNamara Groups). Such areally extensive
sequences often overlie laterally more discontinuous occurrences of bimodal volcanics
and coarse, immature sandstones/quartzites (e.g. Argylla Formation; Haslingden Group;
Carter's Bore Rhyolite; Fiery Creek Volcanics). On the basis of this sedimentary, and
additional geochemical evidence in the western and central parts of the Mount Isa Inlier,
these sequences of volcanics/coarse sediments, overlain by shelf sediments, have been
interpreted as ensialic rift and sag sequences (Glikson et al., 1976; Derrick, 1982; Ellis &
Wyborn, 1984; Sweet, 1985; Blake et al., 1985; Blake, 1986, 1987), deposited during
two (or three) extensional events and their respective tectonic quiescence/thermal
relaxation periods. There are no indications that oceanic crust was formed during any

Table 1.1 Tectonic events of the Mount Isa Inlier. Dates after Page (1978, 1983a)
and Page & Bell (1986).

<table>
<thead>
<tr>
<th>Date</th>
<th>Event Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1780 Ma</td>
<td>Major extensional event (1st &quot;rift + sag&quot; sequence)</td>
</tr>
<tr>
<td>=1750 Ma</td>
<td>Local extensional structures</td>
</tr>
<tr>
<td>1520 Ma</td>
<td>Deposition 2nd &quot;rift + sag&quot; sequence in the west</td>
</tr>
<tr>
<td>1550 Ma</td>
<td>Deposition of first rift and sag sequence</td>
</tr>
<tr>
<td>1450 Ma</td>
<td>Various contractional structures</td>
</tr>
</tbody>
</table>

sage. The first cover sequence consists of the 1860 Ma. old Leichhardt Volcanics,
which are only exposed in the Kalkadoon-Leichhardt Belt. The second cover sequence is
the first rift and sag sequence, deposited during and after De1 (Table 1.1), the first
extensional event of importance here. De1 has been dated by its felsic volcanics at 1780
Ma. (Page, 1978, 1983a). The second extensional event is manifested in the central and
eastern parts of the inlier by local detachment-like structures, extensional duplexes, high-
gle (to bedding) normal faults, all affecting the first cover sequence, and large amounts

\[ \begin{eqnarray*}
\text{Fig. 1.3 Schematic diagram showing the relations between the major stratigraphic,} \\
\text{structural, and magmatic events that affected the Mount Isa Inlier (after Wyborn et al., in} \\
\text{press). Geochronological information is from Page (1983a, 1983b), Page & Bell (1986) and} \\
\text{Wyborn & Page (1983a). The position of the Soldiers Cap Group is contentious,} \\
\text{and, as will be argued in chapter 2, could well be a time-equivalent of cover sequence 2.} \\
\text{Also, in chapter 2 (and in appendix 1), the evidence for a third cover sequence in the} \\
\text{eastern Fold Belt (and Kalkadoon Leichhardt Belt) will be assessed, and it will be} \\
\text{suggested that (parts of) the Mount Albert Group might in fact be part of the Mary} \\
\text{Kathleen Group.} \\
\end{eqnarray*} \]
Note: overall probably approx. 35%; locally 55%.

Holcombe & Fraser (1973) mention 65-80% shortening for the Mary Kathleen Fault Belt.

The important point is that shortening (as in most Hercyno-type orogens) is less than in Alpino-type orogens.
of metamorphosed dolerite dykes and "A"- and "I"-type granites (Passchier, 1986a; Holcombe et al., 1987; Pearson et al., 1987; Passchier & Williams, in prep.; Oliver et al., in prep.; Loosveld, Chapter 2). Its age is estimated at 1750 Ma., as it affects the sag sequence of De1 (Passchier, 1986a; Passchier & Williams, in prep.; Loosveld, Chapter 2), but does not affect 1740 Ma. old granites in the zone of extensional De2 structures (R.W. Page, pers. comm., 1988). The third extensional event (De3) is inferred by an unconformity and another, 1670-1680 Ma. old, "rift and sag" sequence in the western part of the inlier.

It is important to realize that, since the sag deposits, resulting from thermal relaxation, of both extensional events are thick (in the order of 5 km) and are pre-metamorphic, there cannot be a direct link between these extensional events and the anomalously high temperatures during the prograde metamorphism (which is thought to have accompanied regional shortening between 1600 and 1550 Ma.).

The first compressional event to affect the cover sequences, De1, was an, as yet only locally recognized, thrusting event, resulting in imbricate stacks of thrust sheets (Bell, 1983; Loosveld & Schreurs, Appendix 1), fold nappes and bedding-parallel LS-fabrics (Loosveld, Chapter 3). Correlation of the De1 structures from one area to the other is hampered considerably by the different movement vectors that have been proposed for the different areas. The subsequent De2 event was a phase of strong E-W directed coaxial shortening, resulting in upright, tight, N-S trending folds, vertical axial-plane foliations, and vertical extension lineations. De2-shortening is penetrative over the inlier and amounts to 35-55%. Approximating De2 by plane strain, it must have thickened the crust by a factor of between 1.5 and 2.2. In the numerical simulations (Loosveld, Chapter 4), I have assumed crustal thickening by a factor 1.6 or 2. Post-De2 deformation is in the central and western parts in the inlier generally characterized by steep faults and shear zones, trending 030°N (dextral) and 310°N to 340°N (sinistral), with a dominant strike-slip offset, and reflecting waning E-W compression. Some of these faults have been reactivated until the Tertiary. Specific parts of the inlier are affected by late N-S shortening (e.g. the central eastern part; Chapter 2), and NNW-trending, upright folds (e.g. north of Mt. Isa township; Bell, 1983), and other non-pervasive local structures.

1.3.2 Geochronology

As stated, De1 is reliably dated at 1780 Ma. (Page, 1978, 1983a). De2, the second extension, poses more problems, but must postdate 1760 Ma. and predate 1740 Ma. (R.W. Page, pers. comm., 1988; Loosveld, Chapter 2). De3 has been dated at 1680-1670 Ma. (Page, 1978). One De1 shear zone in the Sybella Granite in the west of the
inlier has been dated, giving an age of 1610±13 Ma. (Page & Bell, 1986). However, $D_{c1}$ is only recognized in isolated areas, and correlation of it over areas in between has as yet been ambiguous. Hill (1987; in prep.) studied the structural setting of two U-Pb zircon dates in the eastern part of the inlier, originally interpreted by Page (1983a) as a 1603 Ma. old rhyolite and a 1600 to 1624 Ma. old microgranite, and reinterpreted them as older felsic volcanic rocks recrystallised and deformed during $D_{c1}$, thus confirming Page & Bell's age of $D_{c1}$. Page & Bell (1986) dated a $D_{c2}$ shear zone in the Sybella Granite at 1544±12 Ma.. U-Pb systems of uraninite in the Mary Kathleen uranium deposit were initiated synchronous to the regional metamorphism at 1550±15 Ma. (Page, 1983b). Studies of porphyroblast textures generally confirm the synchronicity of $D_{c2}$ and the regional metamorphism. $D_{c3}$ is dated at 1510±13 Ma. (Page & Bell, 1986). A large population of K-Ar dates of 1450-1500 Ma. (Richards et al., 1963) is interpreted as reflecting a widespread cooling of the crust and cessation of Ar diffusion during uplift.

Major I-type granite suites intruded the upper crust at 1860, 1800, 1740-1720, 1670 and 1500 Ma. (Page, 1978; Nisbet et al., 1983; review by Blake, 1987).

1.3.3 Petrology

Regional prograde metamorphism was of low-pressure facies ($=low P/T$ ratios; Miyashiro, 1973), manifested by the andalusite-sillimanite facies series. Peak-metamorphic assemblages attained are around 400MPa (4 kbar) and 600 to 680°C in the Soldiers Cap Group (Jaques et al., 1982), whereas similar conditions, 300 to 400MPa at temperatures between 560 and 630°C, were inferred for the Mary Kathleen Fold Belt (Wonga-Duchess Belt; Fig. 1.2) by Oliver & Wall (1987) and Oliver et al. (in prep.). Within the Mary Kathleen Fold Belt, the Mary Kathleen Syncline yielded 350 to 480MPa at temperatures between 600 and 650°C (Derrick, 1980), and 350±100MPa at 530-560°C (Hamilton, 1985). Calculations of geothermal gradients in the Mary Kathleen Sheet area by Derrick et al. (1977c), using Holdaway's (1971) determination of the alumino-silicate triplepoint, yielded 35 to 40°C/km for the Blockade Block (20 km NW of Mary Kathleen "township") and 48 to 57°C/km for the eastern succession. In the western part of the inlier, Hill et al. (1975) estimated a metamorphic pressure of 400MPa and a temperature of 650°C for an area close to the Sybella Granite. Cordierite/biotite and sillimanite schists occur e.g. east of the Sybella granite (Wilson, 1973), consistent with the high thermal gradients. Over the entire inlier, therefore, paleo-geothermam gradients in the upper 12-15 kilometers range between 34 and 57°C/km with an average of 44°C/km.

In the inlier, the peak of prograde metamorphism is generally thought to be synchronous with $D_{c2}$, i.e. with the E-W compression (Wilson, 1973; Jaques et al.,
1: Introduction

1982; Passchier, 1986b; Reinhardt & Hamilton, in press; Loosveld, Chapters 2 & 3; M.J. Rubenach, pers. comm., 1988). In the Soldiers Cap Group, it certainly postdates Dc1, as the isograd pattern cuts the Dc1 fold nappe structure. It is probably coeval with early-Dc2, as slightly curved inclusion patterns (S1) in andalusite porphyroblasts contrast strongly with the external foliation, which is commonly tightly crenulated around S2 (Loosveld, Chapter 3). Since metamorphism took place during thickening of the crust, the early, steep thermal gradients disagree strongly with the classical concept of early, high-P facies metamorphism, as e.g. the Eo-Alpine metamorphism in the European Alps.

The control on the metamorphic history in the Mount Isa Inlier is still weak. However, the few P-T-t data that do exist indicate a poly-metamorphic, anti-clockwise path: sillimanite grows usually after andalusite, whereas kyanite replaces sillimanite and cordierite in late retrograde shear zones (Reinhardt & Hamilton, in press). Similar metamorphic observations have been made in other Australian Early to Middle Proterozoic inliers, especially in the Willyama (Phillips & Wall, 1981; Hobbs et al., 1984; Clarke et al., 1987), Halls Creek (Gemüts, 1971), and Arunta (Warren, 1983) mobile belts (Fig. 1.1): the prograde low-P facies metamorphism in all cases is succeeded by a phase of isobaric (or slightly compressive or decompressive) cooling.

The post-tectonic, essentially isobaric cooling path, as deduced from metamorphic data, is confirmed by the sedimentary record. In places, up to 15 km must have been eroded, but large amounts of flysch and molasse type sediments are absent in and around the inlier, possibly indicating the absence of fast uplift rates nearby. Only the lower Soldiers Cap Group resembles a classical flysch deposit, while the Quamby Conglomerate (in a fault-bounded syncline 20 km north of Cloncurry township) is the only possible candidate for a molasse deposit. The Soldiers Cap Group, however, predates deformation and metamorphism. The Quamby Conglomerate is volumetrically insignificant, and it too predates some of the E-W-directed shortening. In short, there is a gap in the chrono-stratigraphic record with the Phanerozoic Georgina, Carpentaria and Eromanga Basins overlying the Middle Proterozoic Mount Isa Inlier. Apatite fission track ages of 337±60 and 240±23 Ma. (Page, 1978, p. 159) indicate an uplift of minimally 5 km between the Late Paleozoic/Early Mesozoic and present.

For the greater part, the above-outlined framework is based on data from the western part of the inlier. Extensive reference lists have been published by Carter et al. (1961), Derrick et al. (1971) and Blake (1987). My study has concentrated on the MARY KATHLEEN, MARRABA, CLONCURRY and KURIDALA 1:100 000 Sheet areas (Fig. 1.2; map commentaries respectively by Derrick et al., 1977c; Derrick, 1980; Ryburn et al., in prep.; and Donchak et al., 1983), in the central eastern part of the inlier.
1.4 ORGANISATION AND FORMAT OF THE THESIS

The body of this thesis consists of 3 papers, two accepted for publication (Chapters 3 & 4), and one submitted for publication (Chapter 2). In chapter 2, an overview of the lithological/structural architecture of the central eastern Mount Isa Inlier is presented, with special emphasis on the *basin-forming* tectonics. It is concluded, in agreement with Wyborn & Blake (1982), that the early evolution of the inlier appears to be entirely ensialic. In chapter 3, the *basin-modifying* structures of one particular area, the central Soldiers Cap Group, are described. In it, evidence is supplied for the contemporaneity of low-P facies metamorphism and crustal thickening. This, in strong contrast to the high-P facies metamorphism in subduction zones, is indicative of a compression mechanism other than that at a destructive continent/ocean plate margin. Chapter 4 discusses the various models which could possibly explain the contemporaneity of low-P facies metamorphism and crustal thickening. The problem is quantitatively addressed: the results of a one-dimensional, finite-difference modelling study of heat transfer in a tectonically active continental lithosphere are presented. Chapter 5, finally, is a brief synthesis of the tectono-thermal evolution of the eastern Mount Isa Inlier. Nine appendices accompany this thesis, the last one in the form of a map. The first appendix, which is directly reproduced from its publication format (from: "Australian J. Earth Sci.", co-authored by Guido Schreurs), describes early compressional structures, and, as will become clear, is thus supportive of the model proposed in chapter 4 and appendix 8. Appendices 2 to 7 contain the technical details of the numerical modelling study. Appendix 8 (co-authored by Mike Etheridge) gives the wider implications of this modelling study. Appendix 9, a 1:100 000 scale map of the central Soldiers Cap Group belt, is in the form of a map in the pocket in the back. Transparent sample location maps, to be overlain respectively on the MARRABA, CLONCURRY and KURIDALA 1:100 000 Sheets (BMR) are lodged with the samples at the Museum, Department of Geology, Australian National University, Canberra, A.C.T. 2601, Australia.

To improve readability, all manuscripts (except the paper in appendix 1) have been adjusted to a thesis format. Repetitions, e.g. in the respective introductory geological framework, have been avoided as much as possible. For the same purpose, references are grouped together at the end. Grid reference numbers refer to the BMR 1:100 000 Sheets. I have also tried to avoid referring to unpublished work, but with thoughts on the evolution of the inlier at present in a fast flux, this was not always possible, specifically in the case of three key manuscript references, namely Ryburn et al., Oliver *et al.*, and Passchier & Williams, all of which are (about to be) submitted for publication. Abbreviations of fabric elements (e.g. S₂, L₁) are after Bell & Duncan (1978).
The reader is strongly advised to keep the MARY KATHLEEN, MARRABA, CLONCURRY, and KURIDALA 1:100 000 Sheets at hand. A generalized flow chart of the numerical experiments, used in chapter 4 and appendix 8, is in appendix 6, as well as 4 programme examples. A 3.5" 2S2D (800K) floppy disc, containing the total programme library, is available from the author on request.
CHAPTER 2

TECTONIC EVOLUTION OF THE EASTERN MOUNT ISA INLIER (MARRABA, CLONCURRY), NW QUEENSLAND, AUSTRALIA

LOOSVELD

In modified form submitted to "Precambrian Research"
TECTONIC EVOLUTION OF THE EASTERN MOUNT ISA INLIER (MARRABA, CLONCURRY), NW QUEENSLAND, AUSTRALIA.

2.1 INTRODUCTION

In the previous chapter the fixist and mobilist viewpoints were introduced, and arguments for both briefly discussed. It was pointed out that the mobilist viewpoint for the Mount Isa Inlier is based on four lines of evidence: 1) the early geochemical database, in which an eastward change from continental tholeiites to ocean floor tholeiites was recognized (Glikson et al., 1976); 2) the geochronology of rocks east of the inlier, which gave consistently younger dates than those of the inlier; 3) the restriction of deep-water sediments to the eastern part of the inlier; and 4) the crustal thinning in the eastern part of the inlier, postulated on the basis of higher metamorphic grades and higher ductility. As to point 1, Wyborn & Blake (1982), on the basis of a greatly enlarged geochemical database, refuted the possibility of formation of oceanic crust anywhere in the inlier. The second argument was also refuted by Wyborn & Blake (1982), as oceanic crustal material has not been encountered in drillholes in the Eromanga Basin east of the inlier, but, in contrast, Precambrian granites and metamorphics do occur east of the inlier. Tentative correlation schemes between the Mount Isa Inlier and the Georgetown Block (Fig. 1.1) have now been proposed (Laing & Beardsmore, 1986).

Points 3 and 4 relate specifically to the eastern part of the inlier. Unfortunately, it is this eastern part, in which the structural framework is the least well understood/ documented and the geochronological control is the poorest. The only major turbiditic sequence does occur in the eastern part of the inlier, but no sedimentary study has so far confirmed the deep-water origin of the turbidites. On the other hand, the pelitic turbiditic sequence has attracted some detailed metamorphic studies (Jaques et al., 1982), in which the low-P facies, near-anatectic metamorphism of the lower sequence was described. This does not mean, however, that such high-grade metamorphism did not affect other parts of the inlier as well, and in fact such metamorphic grades have also been documented for isolated areas in the central and western parts of the inlier (Wilson, 1973; Hill et al., 1975; Derrick et al., 1977c; Oliver et al., in prep.). The aim of my field project was to clarify some of the structural and stratigraphic relations in this eastern part of the inlier, and thus to obtain a more tightly constrained paleo-geographic reconstruction.

The eastern part of the Mount Isa Inlier is here defined as that part of the inlier which lies east of the Pilgrim Fault or its northern extension, the Ballara-Corella River Fault Zone (Fig. 2.1; the Quamby-Malbon and Cloncurry-Selwyn Zones of Blake,
1987). This eastern part strongly differs in three ways from the more western parts. Firstly, the litho-stratigraphy, although in places crudely resembling parts of the central sequences, is in detail very different from its more western counterparts. Secondly, the structural grain deviates from the consistently N-S trending grain in the central and western parts: a NE-trending anticline dominates a large portion of this eastern part, and E-W trending structures typify an E-W strip 20 km wide in the CLONCURRY 1:100 000 Sheet and tapering out towards the west. The occurrence of inverted strata is common in the east, but rare in the other parts. Late faulting is more complex in the east, with no obvious NNE-trending dextral faults and NW-trending sinistral faults. Reverse displacements along steeply ENE- to E-dipping faults, on the other hand, are common in the east. Thirdly, the eastern part is characterized by large batholiths which postdate both the penetrative deformational event and the peak of metamorphism. Such post-metamorphic granites are absent in the rest of the inlier. Pre-metamorphic granites, on the other hand, are abundant in the central and western parts of the inlier, whereas only volumetrically small micro-granites predate metamorphism and the main deformation in the eastern part (Wyborn et al., in press). It must be concluded that the understanding of the geometry and dynamics of the Pilgrim Fault and Ballara-Corella River Fault Zone is of major importance for the reconstruction of the development of the inlier, and that the fault zone might separate two (only weakly-related) terranes. Extreme care has to be taken with the correlation of litho-stratigraphic or structural features over the fault zone.

In contrast to the central and western parts of the Proterozoic Mount Isa Inlier, which are characterized by roughly N-S trending belts, in the central eastern part, i.e. east of the Ballara-Corella River Fault Zone, three major tectono-stratigraphic blocks rather than belts can be recognized (Fig. 2.1). Each block is characterized by a different lithological sequence and a different style of deformation. The blocks are, from west to east: 1) the Mitakoodi Anticlinoirium, a tight, NE-plunging, composite anticline with a wavelength of 50 km, which is developed within subaerial to shallow-marine, continental-rift facies lithologies; 2) the Marimo-Staveley Block, in which a series of tight to isoclinal folds is developed within a stratigraphically complex marine shelf; 3) the central part of the Soldiers Cap Group belt, which is dominated by a series of meta-

\[\text{ Fig. 2.1 }\quad\text{Map of the central eastern part of the Mount Isa Inlier, showing the Mitakoodi Anticlinoirium (I), the Marimo-Staveley Block (II), and the central Soldiers Cap Group belt (III). The incoherent "block"-like map pattern is due to post-D_{c2} faults, such as the Cloncurry "Overthrust", the Big Mick Fault and the Ballara-Corella River Fault Zone, and to the intrusion of the post-D_{c2} Naraku and Williams Batholiths. AB, CD and EFG are the cross-section lines of Fig. 2.5. Numbers 1 to 7 refer to stratigraphic sections in Fig. 2.2. Boundaries of this map are outlined in Fig. 1.2. Below: Schematic E-W cross-section through the centre of the map.}\]
turbidites, and folds with a wide range of trends. The three blocks occupy large parts of the CLONCURRY, MARRABA and northern KURIDALA 1:100 000 Sheet areas (map commentaries are respectively by Ryburn et al., in prep.; Derrick, 1980; Donchak et al., 1983). The transition from subaerial sediments in the west through marine shelf sediments to turbidites in the east has been one of the main arguments used by advocates for a Proterozoic continental margin at this locality (mobilists' argument 3, above). Stratigraphic correlations between these three blocks, however, are contentious (Derrick et al., 1971; Derrick, 1980; Blake et al., 1983; Donchak et al., 1983; Blake et al., 1984; Ryburn et al., in prep.). Stratigraphic correlation is hampered by rapid lateral facies changes, metasomatic alteration, the scarcity of geochronological data, extensive hydrothermal brecciation, some of which obscures critical boundaries, and, most importantly, a poorly-understood structural framework.

In this chapter, I will describe the structure of the three domains. In most cases, megascopic structures had already been mapped on the existing 1:100 000 BMR Sheets. A comprehensive correlation of structures, however, had not been made. I spent approximately 8 months in these 3 domains (another 4 months in the Mary Kathleen Sheet area; Appendix 1), defining and mapping key areas, and subsequently, tracing fabric elements between these key areas. Microscopic and XRD studies supported and complemented the thus erected structural framework (229 30 |im thin sections are lodged with the accompanying hand specimens at the museum, Department of Geology, ANU; 64 additional 30 |im BMR thin sections, originally cut for June Hill, were also studied). The emphasis here will be on the structural correlation; only brief lithological descriptions are given. Lithological details have been presented in the works cited above. A description of the structures within and between the three blocks precedes a discussion of the implications of these structures for the stratigraphic correlation and for the early tectonic evolution of the eastern margin of the Mount Isa Inlier. E-W cross-sections through the three domains are presented in Fig. 2.5. Unfortunately, the construction of balanced cross-sections is inhibited by factors such as the layer-dependent amount of shortening due to solution/precipitation processes, the inter-layer slip and flow, and the absence of undeformed terranes (hinter- or fore-lands) which should give the original thicknesses of the lithologies. Further complications arise from the fact that the strain is clearly not planar. Also, large areas are occupied by brecciated calcisilicate rocks. Although some foliated breccias are probably due to the brittle response to early (Dc1) deformation (Andrew Tunks, pers. comm., 1987), much of the brecciation clearly postdates the penetrative ductile deformation.

It will be suggested here that Dc2 styles and orientations are more diverse in the eastern part of the inlier than in the central and western parts, and that the three tectono-stratigraphic domains of the eastern Mount Isa Inlier are also to a large extent the result of
10- to 100-km-scale D\textsubscript{c2} structures. In fact, each block hosts a D\textsubscript{c2} structure different in style from that in the adjacent blocks. Thus, the style of the map-scale structure depends on the composition of the original rock sequence, similar as, on a smaller scale, deformation depends on the competency contrast and thickness of beds. On the basis of structural relations and their implications for the stratigraphic correlation, I will in this chapter follow the conclusions of more recent studies (notably Blake, 1986, 1987), and refute the suggestion of Plumb & Derrick (1975), Dunnet (1976b), and Wilson (1978) of an eastward progressively deepening basin representing an Early to Middle Proterozoic continental margin. I will attempt to demonstrate that D\textsubscript{c2} folding and differential uplift along D\textsubscript{c3} fault and shear zones led to the juxtaposition of different stratigraphic levels, and to the apparent deepening of paleo-basins from subaerial in the west to deepwater in the east.

2.2 THE MITAKOODI ANTICLINORIUM

2.2.1 General

The Mitakoodi Anticlinorium, immediately east of the Ballara-Corella River Fault Zone is 100 km long and 50 km wide, and comprises the Duck Creek and Bulonga Anticlines and the Wakeful and Boomerang Waterhole Synclines (Fig. 2.1). The Mitakoodi Anticlinorium is, despite its large size, mainly coherent and preserves a relatively undisturbed stratigraphic sequence (described in detail by Derrick et al., 1971). The cores of the anticlines expose intricately intermixed bimodal felsic meta-volcanics and feldspathic and micaceous meta-sediments of the Argylla Formation (Fig. 2.2), the youngest formation of the Tewinga Group (Derrick et al., 1976a). A crude threefold stratigraphy occurs in the Argylla Formation within both cores, from base to top: 1) a feldspathic to micaceous quartzite, 2) massive, porphyritic rhyolites, dacites and andesites with minor feldspatic quartzites, and 3) a well-bedded medium to coarse quartzitic unit. Basalts occur sporadically. Page (1983a), using the U-Pb Zr method, dated a partly recrystallized porphyritic rhyolite in a sequence of relatively low-grade
felsites from the Argylla Formation (MARRABA G.R. 204853) at 1766±23 Ma., a date which is in good agreement with the U-Pb Zr ages of 1783±5 and 1777±7 Ma., obtained from similar rhyolites in the Argylla Formation to the northwest (respectively PROSPECTOR G.R. 818388, Page, 1983a, and PROSPECTOR G.R. 816391, Page, 1978).

The Argylla Formation is conformably overlain by the Marraba Volcanics, the lower formation of the Malbon Group (Derrick et al., 1971; Derrick et al., 1976b). The Marraba Volcanics in the northern part of the Mitakoodi Anticlinorium can be subdivided into three members, with overall fining-upward sediments, from base to top: the Cone Creek Volcanic Member, a rather homogeneous series of subaerial to shallow marine metabasalts with minor intercalations of feldspathic quartzite, then the Mount Start Member, a laminated calcareous, and in places stromatolitic, sand/siltstone sequence, up to 100m thick, and on top the Timberoo Member, consisting of a variety of fine-grained meta-sediments (Derrick et al., 1971). The Mitakoodi Quartzite, the upper formation of the Malbon Group, conformably overlies the Marraba Volcanics (Fig. 2.2). It is dominated by a medium to coarse feldspathic quartzite. (A) metabasalt(s), known as the Wakeful Metabasalt Member, is (are) intercalated within this quartzite. Towards the top, the quartzite grades via siltstones and slates into the Overhang Jaspilite in the west, north and northeast, and into the Answer Slate in the southeast. These form the base of the Mary Kathleen Group (Derrick et al., 1977a).

The Overhang Jaspilite consists of limestones and argillites with jaspilitic beds, the Answer Slate of laminated meta-siltstones, slates, phyllites and minor meta-arenites. The Overhang Jaspilite is overlain by three different meta-sedimentary sequences: from the north to the southeast respectively the Corella Formation, the Marimo Slate and the Answer Slate. All three are thought to conformably or disconformably overlie the Overhang Jaspilite (Derrick et al., 1971; Blake et al., 1983), and therefore to be correlatives (the transition, however, is not exposed). The top of the Overhang Jaspilite is poorly defined. A quartzitic breccia, known as the Chumvale Breccia, occurs along most of the northern and northeastern Mitakoodi Anticlinorium. It originated from the Overhang Jaspilite by brecciation, decalcification and silicification (Derrick et al., 1971). The Overhang Jaspilite is ubiquitously strongly deformed.

The Overhang Jaspilite and Answer Slate mark the boundaries of the Mitakoodi Anticlinorium. In the west, the anticlinorium is bounded by the Ballara-Corella River Fault Zone, a NNE-trending extension of the Pilgrim Fault; to the north lies the Tommy Creek Block, a higher-grade metamorphic domain of calc-silicate rocks, felsic to intermediate volcanics, and minor marbles, quartzites, and basic volcanics (Hill, 1987),
and to the east the Marimo-Staveley Block. Structural styles in the surrounding domains differ from that in the Mitakoodi Anticlinorium (Fig. 2.1).

2.2.2 Depositional environment

Ripple marks and crossbedding in quartzites of the Argylla Formation indicate a shallow-water environment, whereas the rapid lateral facies changes indicate tectonic activity. The amygdaloidal metabasalts of the Cone Creek Metabasalt Member were deposited in a gradually deepening basin, as pillows occur more frequently towards the top. Pillows have not been observed in the base of the Cone Creek Metabasalt Member (also Derrick et al., 1971). The unit was extensively sampled by me for geochemical investigation by L.A.I. Wyborn (BMR). The fine-grained calcareous to cherty quartzites of the Mount Start Member and Timberoo Member are near-shore to shelf deposits. The strongly crossbedded, sometimes ripple-marked, and fining-upwards Mitakoodi Quartzite represents a near-shore to shallow shelf environment and is indicative of a transgressive shoreline. Cherty, very fine-grained siltstones, finely laminated limestones, evaporites and BIFs of the Overhang Jaspilite were deposited on a shallow shelf, consistently below wave base, during a period of tectonic quiescence.

In general, the sequence from coarse, immature, subaerial clastics and felsic volcanics (Argylla Formation) through more mature sediments and mafic volcanism (Malbon Group) to silts, shales and BIFs (Overhang Jaspilite) represents a deepening basin under progressively quieter tectonic conditions, and is characteristic of a large rift basin. The extensional event during which this rift basin developed is abbreviated here by \( D_{e1} \) (Table 2.1). As there is no marked asymmetry in the Mitakoodi Anticlinorium area, and no major syn-sedimentary, lithology-bounding faults are recognized in the area, the architecture of the basin is not apparent. The respective distributions of the topmost unit of the Argylla Formation, and the Marraba Volcanics, from base to top respectively the Cone Creek Metabasalt Member, the Mount Start Member, and the Timberoo Member, are restricted progressively further to the east, but they all also thicken towards the east. No extensional structures, which might imply the tectonic excision of units, have been found. Derrick et al. (1971) mention a general thickening of units of the Argylla Formation towards the east, and interpreted this gradual change in thickness, as resulting from deposition within an eastwards deepening basin, with increasing instability towards the west. In more recent rift-basin concepts, however, the general thickening of the Argylla Formation and lower Malbon Group could be explained in terms of an asymmetric rift basin, bounded in the east by a west-dipping normal growth fault, with faster subsidence, but not necessarily faster deepening, in the east than in the west. Current directions agree well with this structure: in the Argylla Formation, an east-west current system is suggested by ripple marks, whereas a southwestwards current
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<td>1480-1560</td>
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<td>WONGA MICROGRANITE</td>
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<td>Mt. Isa Group</td>
<td>Mt. Albert Group ??</td>
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<td>Carters Bore Rhyolite</td>
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<td>1745±17</td>
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<td>WONGA BATHOLITH (+downfaulting Tommy Ck. Block)</td>
<td>local (N-S?) extension</td>
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<td>YELDHAM + BIG TOBY GR.</td>
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<tr>
<td>1865±3</td>
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<td>Leichhardt Volcanics + KALKADOON BATHOLITH</td>
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<td>Wyborn &amp; Page (1983a), Etheridge et al. (1987b)</td>
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Table 2.1 Outline of the geological history of the Mount Isa Inlier, with emphasis on the eastern part (phenomena relating to the central and western parts are in italics).
direction is suggested by crossbedding (Derrick et al., 1971). Current ripples and crossbedding in the Timberoo Member and the Mitakoodi Quartzite indicate a more west to northwestwards current direction (Derrick et al., 1971).

2.2.3 Structure

Despite the fact that, unlike the consistently N-S trending Dc2 structures in the central and western parts of the inlier, the Mitakoodi Anticlinorium trends NE-wards in the hinge area, it shows all the hallmarks of a typical regional Dc2 structure ("D1" of Holcombe & Frazer, 1979; Pearson et al., 1987; Lister et al., 1985; Loosveld & Schreurs, Appendix 1). Its parasitic folds are non-cylindrical, its axial-plane foliations omnipresent and commonly penetrative, with a vertical mineral lineation. In the eastern limb, km-scale "transposition" plus hinge thickening of the Mount Start Member is developed, resulting in isolated hills like Mount Start, Mount Finish, Mount Conno and Mount Brownie (respectively MARRABA G.R. 225791, 256802, 264827 and 266835). All the Dc2 fabric elements locally overprint older thrust-related structures (Dc1). The Dc2 structures themselves are cut by the Wimberu Granite, part of the Williams Batholith. On the basis of various $^{207}$Pb/$^{206}$Pb ages and one U/Pb age, Wyborn et al. (in press) argue for a crystallisation age of 1500 to possibly 1560 Ma., thus constraining the upper time-boundary of Dc2. The only other timing criterion comes from correlation on the basis of style and orientation with Dc2 structures in the Sybella Granite in the western part of the inlier, which have been dated by the Rb-Sr whole rock method as 1544±12 Ma. (Page & Bell, 1986).

In the core of the Mitakoodi Anticlinorium, tight to isoclinal, upright, similar, NE-trending folds with a penetrative slaty cleavage in the axial-plane are developed within the Argylla Formation, the Cone Creek Volcanic Member, the Timberoo Member and the upper Mitakoodi Quartzite (Fig. 2.5 section AB). Mostly these folds are non-cylindrical on the scale of a few meters. The Boomerang Waterhole Block (Fig. 2.1), an elongate doubly-plunging syncline, reflects this non-cylindricity on a larger scale. On the slaty cleavage, a vertical mica-quartz mineral lineation is commonly developed. In the more competent units, like the quartzites of the Argylla Formation and the Mitakoodi Quartzite, spaced stylolitic cleavages are developed with rare mesoscopic folds. Only the Mitakoodi Quartzite in the eastern limb of the Duck Creek Anticline is characterized by E-W trending, upright, isoclinal, 1-3 km-scale folds. Locally, in foldhinges (e.g. MARRABA G.R. 113861), laminated siltstones and slates, like the ones in the Mount Start Member, Timberoo Member and upper Mitakoodi Quartzite, display (differential) crenulation cleavages (sub-)parallel to the slaty cleavages in the limbs of these folds.
Fig. 2.3 Extensional structures in the Mitakoodi Quartzite in the eastern limb of Mitakoodi Anticlinorium (Duck Creek Anticline). A: map of the Mitakoodi Quartzite between MARRABA G.R. 310955 and 300827. The Wakeful Metabasalt Member separates the lower from the middle Mitakoodi Quartzite. Dolerites compartmentalize the area. In many of the (northern) segments, bedding is non-parallel to the imaginary enveloping surface. A reverse sense of displacement typifies contacts between the segments. B: envisaged history of the area: the \((D_{c2})\) high-angle normal faults are reactivated by a late stage of N-S shortening \((D_{c3})\).
The structure within the Mitakoodi Anticlinorium becomes progressively more complex towards the limbs, i.e. towards the higher stratigraphic units, as it changes from basically a simple single-fold system (folding both bedding and a slaty cleavage) into an interference system of an early extensional phase and at least three later contractional phases. The extensional structures are developed in both limbs of the Mitakoodi Anticlinorium. For the western limb, Passchier & Williams (in prep.) describe "detachment"-like structures, very much like those documented by Passchier (1986a) for the Alligator Syncline. In the eastern limb on the other hand, a swarm of metamorphosed bedding-(sub-)perpendicular dolerites transects the Mitakoodi Quartzite. Here, extension was accommodated by (a) the intrusion of a swarm of dolerite dykes in these units (Fig. 2.3), and possibly by (b) high-angle normal faulting. Minimum extension factors were obtained by measuring the amount of dolerite in a section along strike. Unfortunately, the thickness of the dolerite dykes had to be estimated in some cases, due to poor exposure. By assuming thicknesses of 10 and 25 metres for the poorly exposed dykes, and by choosing three different map sections, six independent estimates of the amount of extension by dolerite intrusion were obtained (Table 2.2). The mean extension factor by dolerite intrusion is 41%. The amount of extension accommodated by possible normal faulting is not recoverable, as the geometry of the blocks at present is one of high-angle reverse faulting (Fig. 2.3). The extension of 41% is therefore a minimum (β ≥ 1.41).

Table 2.2  Estimations of the minimum β, from the eastern limb of the Mitakoodi Anticlinorium within the Mitakoodi Quartzite between MARRABA G.R. 310995 and 300827.

<table>
<thead>
<tr>
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<th>assumed thickness of the partly covered dolerites</th>
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<td>10m</td>
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<td>lower Mit. Qtzite*</td>
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<tr>
<td>middle Mit. Qtzite*</td>
<td>1.30</td>
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<tr>
<td>combined lower and middle Mit. Qtzite**</td>
<td>1.20</td>
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As yet, the extensional event has not been directly geochronologically dated. Two possible ages may be considered. First, since neither the dolerite swarm nor the high-angle faults affect the upper Mitakoodi Quartzite and overlying units, the extension may be synchronous with the deposition of the middle Mitakoodi Quartzite and extrusion of the Wakeful Metabasalt Member. Second, it could be possible that the absence of dolerites and high-angle faults in the finer-grained upper Mitakoodi Quartzite, Overhang Jaspilite, Marimo Slate and Corella Formation is due to the more ductile behaviour of
these rock types. In that case, the extension would be much younger. Because of the restricted exposure, it could not be established whether the dolerite dykes feed or crosscut the intercalations of the Wakeful Metabasalt Member. Trace element partitioning data of the dolerites and basalts are as yet also too limited, and the variation in composition between the (metasomatized) dykes and the Wakeful Metabasalt Member too subtle, to genetically relate them (Ellis & Wyborn, 1984). Thus, there is no direct evidence for the age of the intrusion of the dolerite swarm: (a) it could represent the waning stages of the 1780 Ma old extensional event (De1), during which the Argylla Formation and lower Malbon Group were deposited, or (b) it could represent a second "syn- to post-Mary Kathleen Group" extensional event, and be dynamically unrelated to De1.

Correlating the extensional structures by their orientation would support the second possibility. Assuming that the dolerites intruded along E-W trending fault planes, the sense of shear was one of hanging wall blocks to the north (Fig. 2.3). A similar sense of movement can be inferred from the northern margin of the Mitakoodi Anticlinorium. South of the Tommy Creek Block (Fig. 2.1), a normal fault (or narrow fault zone) with a large displacement on it is inferred to be reactivated by a (Dc1) thrust (Hill, 1987). Amphibolite-grade Corella Formation (Derrick, 1980; Hill, 1987) has been thrust over the stratigraphically lower, lower-greenschist-grade Overhang Jaspilite and Mitakoodi Quartzite. Since the Corella Formation is younger than the Overhang Jaspilite and the Mitakoodi Quartzite, it must have been downfaulted before the metamorphism and the subsequent juxtaposition of these units. The difference in metamorphic grade must reflect the minimum vertical displacement on this normal fault. Movement on this fault must also have had a component of upper block to the north. On the grounds of a thick upper Corella Formation in the south of the Tommy Creek Block, Hill (1987, in prep.) argued for extension during the deposition of this upper Corella Formation.

Pearson et al. (1987) also deduced an upper plate to the north sense of shear for an extensional event in the Mary Kathleen Fold Belt (Fig. 3.1). Recent geochronological studies on granites in this fold belt constrained this extensional event between 1759±22 (Pearson et al., 1987) and, from a granite which lacked the extensional fabric, =1740 Ma. (R.W. Page, pers. comm., 1988). East-west trending folded veins in the Overhang Jaspilite and Corella Formation are another manifestation of early N-S extension. On the grounds of possible correlation with these extensional structures, it seems likely that the dolerite swarm intruded shortly after deposition of the lower Mary Kathleen Group, and, as the lower Mary Kathleen Group represents the "sag" (thermal relaxation) phase of the 1780 Ma. old extensional event (De1), indeed represents a younger phase of crustal extension (De2).
Two other factors support the syn- to post-Mary Kathleen Group intrusion of the dolerites, the first one being the straight, discrete character of the dykes. If the dolerites had intruded synchronously with the deposition of the middle Mitakoodi Quartzite in "soft" sediments, a more erratic pattern would be expected (large-scale loading structures as described by Needham, 1978, stockworks, etc.). Secondly, there is no evidence for an unconformity anywhere in the Mitakoodi Quartzite. Instead, it is laterally uniform over large distances, indicating a tectonically quiet environment.

Apart from this extensional event, three compressional events have affected the limbs of the Mitakoodi Anticlinorium, D_{c1}, D_{c2} and D_{c3}. D_{c2} has already been described above as the penetrative deformation event. No early (D_{c1}) mesoscopic folds were found, but very strong LS-fabrics, parallel to bedding, are locally developed along the margins of the Mitakoodi Anticlinorium, e.g. in some fold hinges in slaty units of the Mitakoodi Quartzite and Marraba Volcanics. Within the Mitakoodi Anticlinorium, it is not established whether D_{c1} is extensional or contractional in character. However, in the adjacent Tommy Creek Block, these bedding-parallel foliations are thrust-related (D_{1a} of Hill, 1987).

In the Overhang Jaspilite in the eastern limb of the Mitakoodi Anticlinorium, plunging-inclined isoclinal folds (D_{c2}) have been locally refolded by E-W trending parasitic folds (D_{c3}; Fig. 2.4). Intersection lineations (L^0) are also folded. An approximately N-S directed shortening is deduced for this event. More evidence for N-S shortening comes from the reverse faulting in the Mitakoodi Quartzite (Fig. 2.3), and from the gentle, kink-like warp in the eastern limb of the Duck Creek Anticline. The E-W trending km-scale folds in the eastern limb of the Mitakoodi Anticlinorium are D_{c2} folds, probably rotated into that orientation by D_{c3}. Also, as argued below (§2.2.3.1), reactivation of D_{c1} thrusts between the Mitakoodi Anticlinorium and the Tommy Creek Block (Hill, 1987) may have led to juxtaposition of the different metamorphic facies within these blocks. In this northern margin of the Mitakoodi Anticlinorium, axial-planes of D_{c3} folds are generally moderately- to steeply-dipping towards the N or NNE (good examples of type-III fold overprinting patterns at MARRABA G.R. 080993).

Other post-D_{c2} structures are mainly brittle in character. Orientations of faults, however, are not as constant as in the central and eastern parts of the inlier; they range from N-S to E-W. There are no obvious NNE-trending, dextral fault systems and NW-trending, sinistral fault systems, as in the central and eastern parts of the inlier.
Fig. 2.4  $D_{c2}/D_{c3}$ overprint (KURIDALA G.R. 401712).  A: Steep penetrative $D_{c2}$ folds developed within the Overhang Jaspilite. B: Irregular $D_{c3}$ folds obliquely transect the $D_{c2}$ folds. C: Disharmonic folding, giving rise to a wide variation of fold-axis orientations. D: Penetrative character of $S_2$ in $D_{c2}$ folds. E: Equal-area, lower hemisphere stereogram representation of selected fabric elements of the outcrop (stars represent $P_0^3$, crosses represent $P_0^3$. F: Simplified representation of fabric elements from E.

Fig. 2.5 (p.29-30)  Schematic structural cross-section through the central eastern Mount Isa Inlier. The cross-section lines are depicted in Fig. 2.1. AB: the Mitakoodi Anticlinorium. The added stereoplot represents poles to $S_2$ in the eastern limb of the Mitakoodi Anticlinorium, in the Mitakoodi Quartzite (dots), in the Overhang Jaspilite (crosses) and in the Marimo Slate (stars). The three great circles represent the mean $S_2$ orientations (Bingham axial distribution to the poles of the three populations; bold: Mitakoodi Quartzite; normal: Overhang Jaspilite; dashed: Marimo Slate). Clearly, the pattern swings from an E-W trend in the Mitakoodi Quartzite, via a NNE-SSW trend in the Overhang Jaspilite, to a N-S trend in the Marimo Slate. It is not clear whether this is purely a $D_{c2}$ refraction feature or whether post-$D_{c2}$ re-orientation of folds played a role. Strike-slip faults with dextral displacements dominate the fault pattern west of the Mitakoodi Anticlinorium, and may have blockrotated the anticlinorium;

CD: the Marimo-Staveley Block and the northern part of the Soldiers Cap Group (through the Toole Creek Syncline);

EFG: the Soldiers Cap Group in the south of the Cloncurry 1:100 000 Sheet area.

Abbreviations of lithologies as in Fig. 2.2. Vertical and horizontal scales are equal.
MITAKOODI ANTICLINORIUM
MARIMO-STAVELEY BLOCK  |  SOLDIERS CAP GROUP

(high degree of interpretation)

SOLDIERS CAP GROUP
2.2.3.1 Discussion of the "Tommy Creek Block"

Hill (1987, in prep.) argued that the juxtaposition of the Tommy Creek Block and the Mitakoodi Block in their present relative positions occurred during Dc1, at 1600 to 1620 Ma., the Tommy Creek Block being thrust over the Mitakoodi Block along N-dipping reverse faults. Therefore, the high-grade metamorphism should be pre-Dc1, or, at the latest, syn-Dc1, which would be an unusual situation in the Mount Isa Inlier: in most other documented cases, regional high-grade metamorphism is early- to syn-Dc2 (Wilson, 1973; Passchier, 1986b; Reinhardt & Hamilton, in press; Oliver et al., in prep.). Hill (in prep.) admits to "meagre field and microstructural evidence for early deformation and metamorphism", but bases her interpretation on (1) large garnet porphyroblasts which are flattened parallel to S1 and which have strain shadows parallel to the pre- or syn-Dc1 extension lineation, and on (2) clots of fibrolitic sillimanite which occur parallel to S1 and which are crenulated by Dc2. A limited micro-structural study of porphyroblast-foliation relations in Hill's thin sections, however, revealed that most garnets and staurolite overgrow a well-developed, and slightly curved foliation, S1 (e.g. Fig. 3.4B), and that the fibrose sillimanite (BMR thin section 86531470, MARRABA G.R. 190092) overgrows a muscovite fabric which is parallel to S1. The sillimanite fibres themselves, however, are unaltered, only very slightly undulous, and not crenulated. Individual fibres crosscut each other in Dc2 hinges, rather than being folded. Thus, they mimic the crenulated muscovite fabric, and are therefore syn- to post-Dc2. Therefore, blastesis of the garnet/staurolite/sillimanite assemblage must have been syn- to post-Dc2. The conclusion is inevitable then that most of the upthrust of the Tommy Creek Block took place during Dc3, the phase of regionally limited N-S shortening, along reactivated Dc1 faults. A strong shearing during displacement of the Tommy Creek Block over the Mitakoodi Anticlinorium in Dc3 explains the absence of recognizable Dc2 (N-S trending) structures in the shear zone, it explains the overprinting by Hill's (1987, in prep.) "D1s" folds by "D1b" folds (="re-orientated D2 folds?" in this shear zone, and explains the deformation of large garnet porphyroblasts in this zone. As shown above and below, some families of post-Dc2 structures in other areas of the inlier are also indicative of such late N-S contraction.

2.3 THE MARIMO-STAVELEY BLOCK

2.3.1 General

The Marimo-Staveley Block lies between the northern part of the Mitakoodi Anticlinorium and the Soldiers Cap Group belt (Figs. 2.1 & 2.6). It comprises rocks of the Marimo Slate, Staveley Formation, Roxmere Quartzite, Mick Creek and Toby Barty Sandstone Members, and minor calcsilicate rock units attributed to the Corella Formation
Tectonic evolution

- Roxmere Quartzite (Ppr)
- Toby Barty Sandstone Member (Pky)
- Mick Creek Sandstone Member (Pkk)
- Marimo Slate Quartzite (Pks1)

Various quartzites

Marimo-Staveley Block

Poor exposure

No data (this study)
2: Tectonic evolution

(Fig. 2.2). Lithologies are described by Derrick et al. (1971, 1977a), Donchak et al. (1983), Blake et al. (1979, 1983) and Ryburn et al. (in prep.). All of these lithostratigraphic units have previously been grouped within the Mary Kathleen Group (Derrick et al., 1977a), except for the Roxmere Quartzite which Derrick et al. (1977b) and Derrick (1980) argued is the lateral equivalent of the (younger?) Mount Albert Group (Fig. 1.2). The contact of the Marimo Slate with the underlying Overhang Jaspilite is concordant, but highly strained; considering the close similarities between the upper Overhang Jaspilite and the lower Marimo Slate and the parallelism of bedding and structures in both units, the contact may be conformable.

Structurally, the Marimo-Staveley Block is dominated by tight to isoclinal N-S trending folds. N-S trending ridges of silicified slates commonly outline the structures. In places, however, these zones of silicification transect strata, thus hindering attempts to reconstruct the stratigraphic sequence. Further complications arise from the lateral facies changes, which are abundant and, although often subtle, like transitions from spotted grey slates to non-spotted black slates, sometimes dramatic: high in the sequence are some eight major, unconnected masses of "Roxmere/Mick Creek/Toby Barty quartzites" (Fig. 2.6). In the northwestern part of the block, a semi-continuous, and probably conformable, sequence is exposed (AA' in Fig. 2.6), consisting from bottom to top of:

i: calcareous rocks, both laminated and brecciated, assigned to the Corella Formation on the MARRABA 1:100 000 Sheet;

ii: dominant grey and minor black slates with stromatolitic cherts and minor laminated meta-siltstones towards the top (symbolized Pkm1 on the MARRABA Sheet; Derrick, 1980). The average grainsize increases towards the top of the unit;

iii: a greyish brown to yellow, laminated to thinly bedded, and in many cases brecciated, quartzite, (Pkm2). Intercalated with this quartzite, and towards the top, are increasing amounts of marble, calc-silicate rock and some marl. A few thin bands of vesicular basalt are intercalated with these calcareous rocks (MARRABA G.R. 437984). Cubic casts, in places obviously stratabound, minor heavy mineral laminations, outlining cross-bedding, and minor wave ripples occur in the quartzites;

iv: Roxmere Quartzite: a banded to massive, pink, fine- to medium-grained, wave-rippled feldspathic quartzite, with abundant heavy-mineral laminae, which show (trough) cross-bedding. Most beds contain mm-scale cubic casts, probably after pyrite;

v: bedded and brecciated calc-silicate rocks, limestones and marls. Intercalated in these are some cross-beded quartzites with heavy mineral laminae, which resemble the underlying quartzites. These rocks were mapped as part of the Corella Formation by Derrick et al. (1971).

⇐ Fig. 2.6 Schematic map of the Marimo-Staveley Block. Sections AA', BB' and CC' are discussed in the text. Legend as in Appendix 9.
A similar sequence (BB' in Fig. 2.6), although strongly deformed, can be observed some 20 km southwards. It includes the Toby Barty Sandstone Member instead of the Roxmere Quartzite, but overall the sequence is the same: a pelitic unit, internally coarsening towards the top, is sandwiched between the topmost calcareous member of the Overhang Jaspilite (or Corella Formation) and another calcareous unit. This upper calcareous unit is overlain by a clean, fine-grained feldspathic sandstone/quartzite (the Toby Barty Sandstone Member of Derrick et al., 1971). This quartzite also shows crossbedding, ripple marks and, to a lesser degree than the Roxmere Quartzite in the north, heavy mineral concentrations. Unlike the Roxmere Quartzite, however, it is commonly thinly bedded. At the top, a steeply SW-dipping, deformed conglomerate with a down-dip extension lineation (aspect ratio 8:3:2) is concordantly overlain by intensively silicified, and subsequently boudinaged slates. Considering the anomalously strong deformation, though, the contact between the Toby Barty Sandstone Member and these slates is probably a tectonic one.

At section CC however, in between AA' and BB' (fig. 2.6), the sequence is different, as the Roxmere Quartzite overlies a unit of (from bottom to top) brown, well-banded quartzites, meta-siltstones with few intercalations of calcareous meta-sediments, and marbles. Transitional beds can be recognized between this unit and the Roxmere Quartzite: the brown, well-banded quartzite interfingers with pink, medium-grained quartzite; in both lithologies heavy-mineral laminae become more prominent towards the top. This unit is attributed to the Corella Formation by Derrick (1980; BMR 1:100 000 MARRABA Sheet).

All in all some 8 masses of fine to medium-grained, and generally well-bedded, quartzites - the thickest ~2.7 km thick - crop out at relatively high stratigraphic levels. They have been previously assigned to the Roxmere Quartzite, and the Toby Barty and Mick Creek Sandstone Members (see MARRABA Sheet). The Roxmere Quartzite and the Toby Barty Sandstone Member are probably correlatives (see also Mathews & Woods, 1969; Derrick et al., 1971): the resemblance between the units is strong and both conformably overlie minor banded calcsilicate rock and greyish brown quartzite (Pkm2). Moreover, a third quartzitic unit, the Mick Creek Sandstone Member (Derrick et al., 1971), lies in a similar stratigraphic position and also resembles the Roxmere Quartzite. Like the Toby Barty Sandstone Member, its average grain size increases towards the top: in the lower part some intercalated, laminated meta-siltstones occur, whereas towards the top minor conglomerates are developed. All outcrops of the Mick Creek Sandstone Member on the preliminary CLONCURRY 1:100 000 Sheet (compiled by Hill et al., 1978) have been reassigned to the Roxmere Quartzite on the final version (compiled by Ryburn et al., in prep.).
Due to the lack of recognizable (=traceable) markerbeds within the Marimo Slate, it is not known whether the various quartzitic masses are exact time-correlatives. All 8 quartzitic masses are to some degree fault-bounded. The eastern margin of the northernmost mass at MARRABA G.R. 438990 is illustrative in this respect: it is internally strongly faulted, and locally interfingers with calcisilicate breccias. The breccias are weakly foliated. Although no calcisilicate or slate clasts are found in the Roxmere Quartzite, this margin may well be a syn-sedimentary fault. The interconnection of the various quartzitic masses at the time of deposition must be regarded as doubtful, as syn-sedimentary faults may have bounded more depocentres.

2.3.2 Depositional environment

The existing 1:100 000 scale maps (Derrick, 1980; Donchak et al., 1983) suggest that the Mitakoodi Anticlinorium and the Marimo-Staveley Block are two strongly contrasting structural domains. However, most of the differences can be explained by the contrasting competence and average layering thickness of rock units in the two blocks. No major faults or shear zones separate the Mitakoodi Anticlinorium from the Marimo-Staveley Block and a probably continuous sequence can be traced over the transitional area, from slates and jaspilites to limestones and calcareous siltstones (banded calcisilicate rocks), to (black) slates. Derrick et al. (1971), interpreted an inferred rapid facies change from jaspilite and limestone overlain by siltstone in the west to calcareous siltstone overlain by black slate in the east as a change from a near-shore environment in the west to a deeper-water environment in the east. Although such lateral facies changes are quite common in the Marimo-Staveley Block, this one is questionable: it was based on the correlation of two limbs of an inferred syncline around MARRABA G.R. 370800. The syncline concerned, however, is parasitic on a larger syncline, and in fact the eastern "limb" is stratigraphically higher than the western "limb" (the asymmetry is sketched in Fig. 2.6 at point D). Derrick et al. (1971) have presented similar alternatives (the one favoured here would be that depicted in their figure 67f). The important point is that the conclusion that the depositionary basin was deepening to the east is unjustified, and might be inviting unwarranted extrapolation to include the flysch-like sediments of the Soldiers Cap Group.

The interbedding of siltstones, sandstones and shales of the Marimo Slate with rare ripple marks and graded bedding was interpreted by Derrick (1980) to indicate a subtidal or deltafront environment. The dominance of shales with occasional stromatolites (e.g. MARRABA G.R. 444965) points towards a stable shallow-water environment (photic zone). Coarsening-upwards and the extrusion of a few basalt flows may indicate slight tectonic activity. For the Roxmere Quartzite, a shallow-water basin is suggested by the concentrations of heavy minerals, the ripples and cross-bedding. The Roxmere Quartzite
is probably deposited in (linear?) fault-bounded basins, with subsidence rates equal to sedimentation rates. The occurrence of calcareous sandstones and few marbles towards the top of the unit indicates a decline of the energy-level. R.W. Ryburn (pers. comm., 1986), on the basis of the occurrence of albite/K-feldspar (?) meta-evaporites, argued for the correlation of the Roxmere Quartzite with the Corella Formation (Mary Kathleen Group). As calcareous meta-sediments of the (upper) Corella Formation concordantly overlie the Roxmere Quartzite in section AA' (Fig. 2.6), this suggestion is followed here: reassigning the Roxmere Quartzite from the Mount Albert Group to the Mary Kathleen Group renders the name "Mount Albert Group" obsolete in the eastern part of the Mount Isa Inlier.

A corollary of the interpretation that the Marimo Slate conformably overlies the Overhang Jaspilite is that it supports the interpretation of a rift-related deposition for the Argylla Formation and (lower) Malbon Group. Such rift deposits are commonly overlain by low-energy, areally extensive marine shelf sediments, interpreted to be deposited during subsidence due to thermal contraction (the "sag" phase). The upper Mitakoodi Quartzite, the Overhang Jaspilite and the Marimo Slate could well be such sag deposits. A complete rift→sag sequence within and around the Mitakoodi Anticlinorium can now be recognized: basal, coarse rift sediments and bimodal volcanics (Argylla Formation, Marraba Volcanics), followed by the gradually more steady, low-energy, and areally extensive Mitakoodi Quartzite and Overhang Jaspilite, which in their turn are overlain by a sequence dominated by slates.

In the upper stratigraphical units of the Marimo-Staveley Block, this pattern is disturbed, as coarsening-upwards sequences and rapid lateral facies changes occur. The Roxmere Quartzite and the Mick Creek and Toby Barty Sandstone Members are deposited as thick, in places demonstrably fault-bounded, moderate- to high-energy, crudely rhomb- or wedge-shaped, possibly linear lenses within an overall low-energy environment. Some of the Roxmere Quartzite masses exceed 2 km in thickness, while only 6 km wide along strike. Considering such aspect ratios and the rock types, the depocentres must have been fault-bounded, like a rift or a pull-apart basin (or, in the case of transtension, a combination of these). The occurrence of felsic and minor mafic volcanics in the upper Corella Formation of the Tommy Creek Block, which is a correlative of the upper stratigraphical units of the Marimo-Staveley Block, may be indicative of deep-seated lithospheric extension. As discussed above, extensional Dc2 structures are probably related to the deposition of the Tommy Creek Block lithologies (above; Hill, 1987, in prep.). The scarcity or lack of extrusives in the Marimo-Staveley Block, however, and the rapid lateral facies changes, the relatively thick units, and the local unconformities are criteria previously thought to be diagnostic for strike-slip basins, in which, due to curvatures in the faults, areas of horizontal extension (transtension/pull-
apart basins) are juxtaposed with areas of horizontal shortening (transpression/uplift; Wilcox et al., 1973; Crowell, 1974; Reading, 1980; Rodgers, 1980; Christie-Blick & Biddle, 1985). Possible analogues of the Roxmere Quartzite in classical strike-slip regimes are the 2 km thick fluvial Miocene Hazeva Formation in the Arava Valley in the southern extension of the Dead Sea (Manspeizer, 1985), and the Devonian fans of the Hornelen Basin and other basins between Bergen and Trondheim, Norway (Steel & Gloppen, 1980; Nilsen & McLaughlin, 1985).

It must be emphasized, though, that many of the criteria used to distinguish a volcano-sedimentary sequence deposited in a purely extensional rift from one deposited in a strike-slip basin may in fact be applicable to both these endmembers. Recent studies (notably Lister et al., in press) have emphasized the wide variety of possible rift-basin architectures. The recently recognized transfer faults in passive continental margins (Gibbs, 1984; Bosworth, 1985; Etheridge et al., 1987a; Lister et al., 1986a, in press) introduce strike-slip displacements to rift basins and may well explain fan/drainage area mismatches and depocentre migration with time, previously considered to be the most diagnostic criteria for strike-slip basins (Steel & Gloppen, 1980). Both pull-apart basins and small extensional rift basins can have lateral and longitudinal infilling. Other criteria such as

"...(1) lateral matching of displaced paleogeographies across faults (2) discordance between size and materials of alluvial fans and possible source areas (3) thick, but not laterally extensive, sedimentary piles deposited very rapidly (4) localized uplift and erosion giving rise to unconformities of the same age as thick sedimentary fills nearby (5) extreme lateral facies variations..." (Reading, 1980)

are equally applicable to small rift basins separated by transfer faults as to strike-slip basins. With the present limited understanding of the D_e2 fault geometry and sedimentation versus uplift histories during D_e2 on a much larger scale, discussion about the precise regional state of stress is rather futile. A third possibility that wrench tectonics were superposed on the 1780 Ma. crustal extension, and became predominant after the development of the major infra-rifts, can also not be excluded. As yet, one is limited to the conclusion that D_e2 was, at least locally, an extensional event with an unknown component of strike-slip.

2.3.3 Structure

As in the Mitakoodi Anticlinorium, there is evidence for at least three compressional events. The oldest one is inferred from the fact that bedding in at least 2 of the 8 major masses of Roxmere ("Mick Creek/Toby Barty") Quartzite is inverted, and can safely be assumed to be a thrusting event. In the Roxmere Quartzite around MARRABA G.R. 420010, the inverted beds are folded by an open, N-S trending, upright synform. Additional evidence for an early deformational event comes from the widespread occurrence of an, in places folded, slaty cleavage (S_1) in the fine-grained
sediments. The bedding/slaty cleavage relation at MARRABA G.R. 445959 and 389816 suggests that the Dc1 movement vector had a southerly-directed component. Without an intimate knowledge of the Dc1 and Dc2 fold mechanisms, it is not possible to rotate intersection lineations back to their original orientation, and obtain a more exact movement vector (e.g. shear folding with the sense of shear parallel to the Dc2 extension lineation, L2, as advocated by T. Bell, pers. comm., 1987, would re-orientate the plunge direction of pre-existing lineations in a different fashion than would be the case with flexural flow/slip or "neutral plane" folding). An approximately S-directed sense-of-shear was also deduced for the Dc1 thrusts between the Tommy Creek Block and the Mitakoodi Anticlinorium (Hill, 1987).

The most pervasive structure within the Marimo-Staveley Block is an isoclinal, N-S trending, upright fold system (Fig. 2.7A), centred on a major syncline (Figs. 2.1 & 2.5 section CD). Fold axes are commonly very steep and in places fan through the vertical. In the pelitic rocks, a penetrative slaty cleavage or crenulation cleavage is normally axial planar to these folds. In places, a weak, vertically- to steeply-plunging stretching lineation (both towards the N and S) is developed in the slaty cleavage. Pebbles within the topmost Toby Barty Sandstone Member are vertically prolate. At MARRABA G.R. 340886, the slaty cleavage overprints a tight fold in a finely laminated meta-siltstone. The slaty cleavage is commonly defined by flattened quartz, white micas and flattened spherules. Transposition parallel to the cleavage is commonly intense, leaving only a few recognizable remnants of laminated siltstone (Fig. 2.7B). This deformational event reactivates S1, also a slaty cleavage, on the limbs of its tight to isoclinal, steeply-plunging, N-S trending folds, whereas it crenulates S1 in its fold hinges (e.g. MARRABA G.R. 445959 and 370795).

This second deformational event is, on the basis of style and overprinting relations, attributed to Dc2, the penetrative E-W shortening of the western and central parts of the inlier (Lister et al., 1985; Loosveld & Schreurs, Appendix 1; D1 of Holcombe & Frazer, 1979). With regard to the correlation of structures by means of orientation, it should be noted that the orientation of the km-scale E-W trending folds on the eastern limb of the Mitakoodi Anticlinorium appear to be anomalous as they are perpendicular to the trends in the neighbouring Marimo Slate. However, the E-W trends in the competent Mitakoodi Quartzite gradually swing in the Overhang Jaspilite to N-S trends almost parallel to the folds in the Marimo Slate (Fig. 2.5 stereonet). There is also a continuity of fold styles.

\[ \text{Fig. 2.7 Deformation in the Marimo-Staveley Block: A: folded banded calc-silicate rocks of the Corella Formation (MARRABA G.R. 454013); B: transposition in strongly sheared banded calc-silicate rocks attributed to the Marimo Slate (MARRABA G.R. 398809).} \]
Complete tracing of $S_2$-foliations from the Mitakoodi Anticlinorium into the penetrative slaty cleavage in the Marimo Slate was not possible, due to poor outcrop and the occurrence of silicified breccias along the contact between the Overhang Jaspilite and the Marimo Slate. The third event is sporadically recognizable, and consists of upright to steeply-dipping, E- to ESE-trending crenulations of $S_1/S_2$.

2.4 THE SOLDIERS CAP GROUP BELT

2.4.1 General

The Soldiers Cap Group (SCG), formally defined by Derrick et al. (1976c), occupies the easternmost, NNW-trending belt of the Mount Isa Inlier (Fig. 2.1; Appendix 9). It is one of the most problematical sequences of the inlier as it is both structurally and litho-stratigraphically difficult to correlate with any of the other sequences. In the CLONCURRY 1:100 000 Sheet area, the boundaries of the SCG are inferred to be tectonic, but they are obscured by enigmatic calc-silicate breccias (Glikson & Derrick, 1970; Ryburn et al., in prep.), young faults and young plutons of the 1500 Ma. old Williams and Naraku Batholiths (Nisbet et al., 1983; Page et al., 1984), namely the Saxby Granite in the south and the Naraku Granite in the north (Fig. 2.1).

The SCG has been subdivided into three formations, from base to top: the Llewellyn Creek Formation, the Mount Noma Quartzite and the Toole Creek Volcanics (Fig. 2.2; Derrick et al., 1976c). Boundaries between these formations are transitional. The Llewellyn Creek Formation consists of a monotonous series of cyclothems of well-bedded and alternating graded quartz-intermediate meta-greywackes and meta-pelites, which have been interpreted as meta-turbidites by Glikson & Derrick (1970) and Ryburn et al. (in prep.). Completely developed classic flysch profiles (Bouma, 1962) are common; towards the top base-missing turbidites become dominant (Fig. 2.8). Reworking of the turbidites and amalgamation into massive quartz beds occurs in the formation.

The Mount Noma Quartzite conformably overlies the Llewellyn Creek Formation. Its base is defined by a thick quartzite (Derrick et al., 1976c), whereas the sequence itself consists mainly of base-missing meta-turbidites, cross-bedded and massive, clean

⇐ Fig. 2.8 Meta-turbidites in the Soldiers Cap Group. A: a series in section; B: close-up of porphyroblasts; C: one rotated andalusite porphyroblast; D: top of meta-pelitic layer full of (chiastolitic) andalusite porphyroblasts (A, B, C and D from KURIDALA G.R. 697776); E: pseudomorphic overgrowth of andalusite by sillimanite (CLONCURRY G.R. 855779).
quartzites with, towards the top progressively more abundant, intercalated meta-siltstones, phyllites and chemogenic BIFs. In the upper Mount Noma Quartzite and the Toole Creek Volcanics, basalts become progressively dominant.

The Toole Creek Volcanics conformably overlie the Mount Noma Quartzite, and consist of metabasalts and fine-grained meta-sediments like phyllites, meta-siltstones and cherts. In the Weatherly Creek Syncline area (Fig. 2.1), limestones and carbonaceous slates occur in the Toole Creek Volcanics. Limestones are absent, however, in the same member in the Pumpkin Gully Syncline, 20 km ENE from Cloncurry.

All three formations are intruded by dolerites, ranging in thickness between a few metres and 700 metres. They are mainly sills, but also axial-plane-parallel dykes. All dolerites studied are metamorphosed.

2.4.2 Depositional environment

A few sedimentary structures in the Llewellyn Creek Formation are reminiscent of hummocky cross stratifications (Harms et al., 1975, 1982; Walker, 1982; Walker et al., 1983), and might therefore indicate reworking of the turbidites by storm waves and a depositional environment between storm wave base and effective fair weather wave base, in contrast to the "traditional" deep-water turbidites. Slump structures have not been observed and a high-gradient slope position (e.g. a continental slope adjacent to a feeder channel) can probably be excluded.

In the Mount Noma Quartzite, the overall upward-decreasing grain size and occurrence of BIFs and carbonaceous slates indicates a further deepening of the basin. The general absence of scouring and pebbly greywackes indicates deposition in a moderate- to low-energy environment, e.g. on a lower (or more distal) fan. Syngenetic, stratiform Pb-Zn deposits in Mount Noma meta-sediments of the southern Soldiers Cap Group are possibly due to exhalations from sub-aqueous volcanoes.

The chemical affinity of basalts in the Toole Creek Volcanics has been reported to resemble that of a MORB (Glikson et al., 1976), but re-appraising the igneous geochemical database led Wyborn & Blake (1982) to conclude that they, like the other mafic rocks in the inlier, had a continental tholeiitic signature.

The overall deepening marine basin is confirmed in the southern SCG by Beardsmore et al. (in press). They have also studied the two formations conformably underlying the SCG, and concluded that the SCG and its two underlying members represent an intra-continental rift sequence.
2.4.3 Structure

The structural analysis of the central SCG belt will be the main subject of chapter 3. Here, only a brief outline follows. Interpretations of the structure of the SCG belt differ markedly. Glikson & Derrick (1970) and Ryburn et al. (in prep.) agreed on the overprinting pattern of the Toole Creek and Weatherly Creek Synclines (Fig. 2.1): an "older" Toole Creek/Weatherly Creek Syncline was overprinted by a "younger" NE-trending fold. According to Glikson & Derrick (1970) this younger fold was the Snake Creek Anticline (Fig. 2.1); according to Ryburn et al. (in prep.), however, the Snake Creek Anticline predates the Toole Creek/Weatherly Creek Syncline, whereas the latter is folded around an unnamed NE-trending fold on the northeastern part of the Snake Creek Anticline.

It is argued in chapter 3 that there are three lines of evidence against the above-mentioned models. Firstly, the axial-plane fabric of the Snake Creek Anticline, with a well developed extension lineation in it, is the oldest recognizable fabric and is overprinted by some of the peak-metamorphic assemblages. Because of this, and because of the (sub-)parallelism between isograds and bedding, this fabric, and therefore also the Snake Creek Anticline, must be Dc1 structures. A comparison of the Snake Creek Anticline with well-documented Dc2 structures in the central Mount Isa Inlier shows several differences. Whereas the latter are symmetric and upright, the Snake Creek Anticline is reclined and has a strongly sheared western limb (Fig. 2.5 section EFG). Shear strain in this western limb has gradually increased towards the west. A young, brittle fault, the Cloncurry "Overthrust" (Honman et al., 1939; Carter et al., 1961), has reactivated this western limb; tectonic breccias (and enigmatic young calcsilicate breccias) obscure much of the old shear zone. Erosion of the thrust front (i.e. of the SCG and/or overlying units) accompanying movements along this thrust, might have been a factor leading to the local high T/P ratios as reflected from metamorphic assemblages within the SCG.

Secondly, it appears that the Snake Creek Anticline is of the same age as the E-W trending Toole Creek Syncline. S1 can be traced continuously from minor NNW-trending synforms and antiforms of the Snake Creek Anticline to minor E-W trending folds of the Toole Creek Syncline fold system. It follows that the Toole Creek Syncline is a Dc1 structure too. This conclusion does not necessarily disagree with the observation by Ryburn et al. (in prep.) and B. Skrzeczynski (pers. comm., 1986) that a bedding-parallel slaty cleavage is folded around the E-W trending Toole Creek Syncline folds. As Mitra & Elliott (1980), Sanderson (1982), and Loosveld & Schreurs (Appendix 1) showed, a bedding-parallel cleavage (S1a in the terminology of Loosveld & Schreurs), related to the early stages of progressive nappe development, can be folded by the later
stages of progressive nappe development, and be overprinted by a more upright foliation (S_{1B} or S_{1C}). Also, close examination of these foliations revealed that, commonly they are in fact axial planar to the E-W trending folds, with a strong convergence in the slates giving the false impression of a folded cleavage.

Finally, it will be demonstrated in chapter 3 that the Weatherly Creek Syncline and the Middle Creek Anticline (Fig. 2.9) are younger than the Snake Creek Anticline and the Toole Creek Syncline, since S_1 and its related extension lineation L_1 are folded around them. Thus, the picture of a D_{c1} fold nappe, refolded by D_{c2}, will emerge. As in the central and western Mount Isa Inlier, and in the Mitakoodi Anticlinorium and Marimo-Staveley Block, the major E-W shortening is preceded by a thrusting event. Prograde low-P facies metamorphism, characterized by the sillimanite-andalusite stability field, is syn-D_{c2}. The contemporaneity of D_{c2}, i.e. crustal thickening, and low-P facies metamorphism will be discussed in chapter 4 and appendix 8.

Post-D_{c2} deformation is characterized by moderately to steeply W- to S-dipping crenulation cleavages, 290°N trending lineaments (Appendix 9) with a sinistral sense of shear, and 290°N trending, open to tight folds in the vicinity of the Williams Batholith granites. The Toole Creek Syncline system of E-W folds is overprinted by open, NNW-SSE and NNE-SSW trending folds and crenulations. The NNE-SSW trending crenulations dip dominantly, if not persistently, moderately to steeply to the NW. A steeply east-dipping, high-angle reverse fault, the Cloncurry "Overthrust", separates the amphibolite-grade SCG in the east from the greenschist-grade Mary Kathleen Group in the west. A similar fault in the south, the Mount Dore Fault, separates the Kuridala Formation (correlative of the SCG) from the Staveley Formation (Beardsmore et al., in press). Also, on a smaller scale, reverse displacements along ENE- to E-dipping faults are common.

2.5 REGIONAL STRATIGRAPHY

The tracing of structures from one block to the other leads to the necessity to revise the stratigraphic sequence (Figs. 2.2 & 2.10). It is suggested here that the Marimo Slate conformably overlies the Overhang Jasplilite and is a correlative of the lower Corella Formation, as also suggested by Derrick (1980). A shallow-marine arenitic sequence,

\[ \text{Fig. 2.9 Deformation in the Soldiers Cap Group. A: W-verging D_{c1} folds in the Middle Creek Anticline (CLONCURRY G.R. 851795); B: penetrative NW-dipping D_{c2} crenulations (CLONCURRY G.R. 872814); C: microphoto under crossed nicols of crenulation from B. The (recrystallised) quartz enrichment in the hinges of these crenulations forms a differentiated cleavage.} \]
Fig. 2.10  Schematic depositional history of the central eastern part of the Mount Isa Inlier (east of the Pilgrim Fault/Ballara Corella River Fault Zone). Sense of asymmetry of the infra-rift basins is generally not known, except for the "Argylla/lower Malbon Group" rift in the Mitakoodi Anticlinorium area, which is probably bounded in the east by a west-dipping normal fault (system).
laterally discontinuous, and containing the Roxmere Quartzite and the Mick Creek and
Toby Barty Sandstone Members, forms the top of the Marimo Slate, and is overlain by
the upper Corella Formation (Fig. 2.10). It may be the lateral equivalent of the (upper)
Tommy Creek Block lithologies. Towards the west, along the Corella-Ballara River
Fault Zone, the Corella Formation overlies the Mitakoodi Quartzite and its western
equivalent the Ballara Quartzite. Thus, the Overhang Jaspilite, the Marimo Slate and the
various quartzite formations would be lateral facies equivalents of the lower (and middle)
Corella Formation.

Many correlation schemes between the SCG and other units to the west have been
proposed. Carter et al. (1961), Plumb & Derrick (1975) and Derrick et al. (1976c)
tentatively correlated the SCG with cover sequences like the Malbon (and Haslingden)
Groups (cover sequence 2; Fig. 1.3); with Derrick et al. (1976c) mentioning the possible
correlation of the lower SCG with the Argylla Formation (~1780 Ma.; also cover
sequence 2). Donchak et al. (1983), Blake et al. (1983, 1984), Laing & Beardsmore
(1986) and Beardsmore et al. (in press), on the other hand, correlate it with the Kuridala
Formation, which, according to Beardsmore et al. (in press), unconformably underlies
the Staveley Formation of the Mary Kathleen Group. Beardsmore et al. (in press) have
proposed to group the SCG and its two underlying units, and the Kuridala Formation
into the "Selwyn Supergroup" ("Maronan Supergroup" in Beardsmore et al., 1988), and
interpreted this supergroup as part of an ensialic rift sequence. The supergroup,
according to Beardsmore et al. (in press), is the eastern equivalent of cover sequence 1,
the oldest cover sequence of the inlier, which is geochronologically constrained between
1880 and 1810 Ma. (Blake, 1987; Fig. 1.3). The following nine arguments have been
used by Beardsmore et al. (in press) to support this correlation of the Maronan/Selwyn
Supergroup with cover sequence 1 (my comments to these 9 points will follow at the end
of this list):

1. metamorphic grades are higher in the Maronan/Selwyn Supergroup than in the
adjacent rock units which are attributed to cover sequence 2;

2. the substantial thickness of sediments in the Supergroup (at least 10 km) is
comparable to that of the Leichhardt River Fault Trough (LRFT; cover sequence 2),
while these basal sequences are only ~100 km apart. It is considered unlikely for
two 10 km thick rift sequences to develop synchronously within such a short
distance;

3. the regional disconformity separating the Leichhardt Volcanics (1875-1850 Ma.) from
the Argylla Formation (1780 Ma.) is supposed to reflect an erosional period
(Bultitude et al., 1977; Blake, 1987), but no sediments had been reported which
might have been derived from the Leichhardt Volcanics;
2: Tectonic evolution

(4) the Leichhardt Volcanics are postulated to correspond to the (pre-)rift phase of cover sequence 1 (Blake et al., 1985). No "sag" packages, however, are recognized in cover sequence 1;

(5) the SCG is said to be unconformably overlain by the Mary Kathleen Group, whereas in a rift basin, rift and sag sequences should be roughly conformable or disconformable;

(6) while volcanics of the supergroup, like those of cover sequence 1 (1860 Ma. old Leichhardt Volcanics; Fig. 1.3), are generally weakly magnetic or non-magnetic, those of the Argylla Formation and Marraba Volcanics (cover sequence 2) are magnetic. The difference in magnetic signatures may reflect different sources and thus different source ages;

(7) the sedimentary signature differs too: the supergroup contains the only (meta-)turbidites in the inlier, as well as the only plagioclase-rich (meta-)sediments, believed to be extrusive;

(8) the presence of BIF-hosted Pb+Zn±Cu mineralization is unknown elsewhere in the inlier;

(9) the close similarities between the Maronan/Selwyn Supergroup and the Willyama Supergroup (Willyama/Broken Hill Province; Fig. 1.1) indicate a similar age. The Willyama Supergroup has been dated at 1820±60 Ma. by the Rb-Sr whole rock technique (Shaw, 1968), whereas Beardsmore & Laing also mention a minimum age of 1800 Ma. from preliminary U-Pb Zr age determinations.

None of these arguments, however, are conclusive. The following 9 points refer to their counterparts above:

(1) amphibolite-grade and sillimanite/K-feldspar grade metamorphics occur also in cover sequences 2 and 3 in the western parts of the inlier;

(2) as the sedimentary (onlap/offlap) and tectonic structures (particularly thrusting) within the cover sequences are not fully understood, the thicknesses of the rift sequences are crude estimates. However, even when bearing this uncertainty in mind, the actual "rift" packages of the Leichhardt River Fault Trough (LRFT) and the Maronan/Selwyn Supergroup are not 10, but probably respectively ~8 (Derrick, 1982) and ~6-10 km thick; the remainder, respectively 2 and 0-4 km, reflects the more uniform thermal subsidence ("sag"). A reconstruction of the Mount Isa Inlier before the E-W shortening increases the distance between the Leichhardt River Fault Trough and the Maronan/Selwyn Supergroup to ~150 km. Beardsmore et al.'s (in press) arguments that the width (l) of a rift (=half graben?) is generally proportional to the amount of rift subsidence (h), and that both the Leichhardt River Fault Trough and the Maronan/Selwyn rift contain thick rift successions, are correct, but the minimum width of a 10 km thick rift basin is for all reasonable parameters smaller than 150 km (Figs. 2.11A & B), i.e. l/h<15, except for very large extension factors or for very
Fig. 2.11 Geometrical relations of a half graben. A: asymmetrical rift, with extension occurring on the "domino"-principle. The thickness (h) of the rift depends on the width (l₁) of the rift basin, and on the dips of bedding (α) and fault (γ), as \( l₁/h = \cot \alpha + \cot \gamma \) (restriction: \( \alpha + \gamma < 90° \)); B: graphical representation of the above relation. The stippled area represents the domain for which \( l₁/h < 15 \) (see text for details); C: graphical representation (from Gibbs, 1986) of the extension factor as a function of α and γ. The condition \( l₁/h < 15 \) is only valid for extremely high extension factors (>500%; Fig. 2.11C), or for extremely shallow faults (<20°). Listricity of the fault would further decrease the width/thickness ratio. It thus follows that the Leichhardt River Fault Trough and the Maronan/Selwyn rift basin are not too close to each other to refute synchronity of the two basins.

shallowly dipping initial normal faults. An instructive example may come from a seismic section through an approximately 8 km deep lower Cretaceous half graben in the Bass Basin, southeast Australia, which shows a dip of bedding, α, of 34° and a dip of the planar normal fault, γ, also of 34° (Etheridge et al., 1987a, their figure 3), and thus \( l₁/h \) equals only 3. Thus, it is very well possible that the LRFT and the Maronan/Selwyn rift basin developed synchronously, and adjacent to each other;

(3) according to Wyborn & Blake (1982), the grain size and feldspar content of the turbiditic meta-sediments in the southeasternmost part of the inlier (both those of the Kuridala Formation and of the SCG) decrease westwards, indicating a source towards the east. This would exclude the Leichhardt Volcanics, which are in the central part of the inlier, as source;
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(4) this argument stems from the hypothesis that both cover sequences 1 and 2 are rift initiated. While the second cover sequence may well be rift-related, evidence for a rift-initiated first cover sequence is very scarce. In fact, the only unit whose age fits the model age of cover sequence 1 (1875-1800 Ma.) is the Leichhardt Volcanics, a series of felsic ignimbritic deposits and presumably subaerial lava flows. The Leichhardt Volcanics are comagmatic with the Kalkadoon intrusive suite (Wyborn & Page, 1983a), which is interpreted as a post-tectonic (i.e. post-compressional!) suite (Wyborn et al., in press);

(5) the contact between the SCG and the Mary Kathleen Group is mostly tectonic (reverse faults, like the Dc3 Mount Dore Fault, the Dc3 Cloncurry "Overthrust" and the Dc1 Pumpkin Gully Thrust; Fig. 2.1 and Appendix 9). Brecciated, and possibly remobilised (Ryburn et al., in prep.), calcisilicates of the Mary Kathleen Group further obscure the contact. In the more southern area around the Starra Au-Cu deposit, Davidson et al. (in prep.) document a "seemingly conformable" contact between the Kuridala Formation (the western equivalent of the Soldiers Cap Group?) and the overlying Staveley Formation of the Mary Kathleen Group;

(6) the wide variation of magmatism in modern rifts (e.g. Riecker, 1979) is instructive here, as it indicates different mantle sources, different chemical conditions, and different amounts of crustal contamination all within one rift zone. All three influence the fO2 and SiO2 content, and therefore the magnetite concentration and magnetic susceptibility of the volcanics (O2 + 3Fe2Si4 <-> 2Fe3O4 + 3SiO2);

(7) as the source of the meta-sediments of the Maronan/Selwyn Supergroup may well have been to the east, it is to be expected that the sedimentary signature of the Supergroup differs from that of more eastern units. As to the occurrence of turbidites: the generalisation that if they do not occur in one rift basin, they should also not be present in the adjacent rift depression, is not warranted. The wide variety of possible basin architectures in an extensional setting should not be underestimated. Such architecture depends on factors as the extent and timing of the local thermal disturbance, the spacing of normal and transfer faults and size of the rift depressions (which depend among other parameters on pre-existing structures), on extensional strain rate, speed of isostatic rebound, the presence of an elevated hinterland, the amount of underplating and volcanism, etc. (e.g. Lister et al., in press, their figure 22). One may point out the many differences between the western (mainly Haslingden Group and Quilalar Formation; Fig. 1.2) and central (Magna Lynn Metabasalt, Argylla Formation and Mary Kathleen Group) cover sequence 2 contents;

(8) as with 7, it is not doubted that the Maronan/Selwyn Supergroup developed in a distinctly different basin from that of the other units of the inlier. What is doubted, though, is that such largely different basins cannot develop synchronously during an extensional event;
According to Blake et al. (1983, 1984), the Kuridala Formation is part of the Mary Kathleen Group. Their correlation of (part of) the SCG with the Kuridala Formation, and thus with the Mary Kathleen Group, is further supported by the fact that the SCG is conformably overlain by, and locally interfingers with, the Doherty Formation (Mary Kathleen Group). The fact that over the entire eastern part of the inlier (>200 km) no volcano-sedimentary unit lies between the SCG and the Mary Kathleen Group also strongly points towards a deposition approximately synchronous with the deposition of the Argylla Formation and Malbon Group (Fig. 2.10). A partly recrystallised porphyritic rhyolite from the Doherty Formation has been dated at 1720±7 Ma. (Page, 1983a). Such a young age for the SCG is doubtful: the early high-grade metamorphism of the SCG and the juxtaposition of the SCG and the Marimo-Staveley Block by upfaulting of the SCG seem to point towards an age for the SCG older than that of any of the units exposed in the Marimo-Staveley Block. The Marimo Slate and Staveley Formation are time-correlatives of the Corella Formation, which is intruded by the 1745±15 Ma. old Burstall Granite (Page, 1983b). A minimum age for the SCG is therefore 1745±15 Ma.. Ryburn et al. (in prep.) correlate the Mitakoodi Quartzite with the Mount Noma Quartzite, whereas, in contrast, Derrick et al. (1976c) correlate the complete Malbon Group with the Toole Creek Volcanics (Fig. 2.2). Both are possible. Although none of the above-mentioned arguments is conclusive, the correlation of the SCG with other units of cover sequence 2 is preferred here. An age of 1780-1750 Ma. seems most likely for the deposition of the SCG.

2.6 DISCUSSION AND CONCLUSIONS

In the last five years, the view that the northern Australian Early to Middle Proterozoic mobile belts were generated in intra-continental rift settings has become increasingly popular with Australian geologists (review by Etheridge et al., 1987b). In the Mount Isa Inlier, evidence for intra-continental rifting was initially based on lithofacies in the Western Fold Belt (Glikson et al., 1976; Derrick, 1982). Blake et al. (1985) and Blake (1986, 1987) considered the history of the Mount Isa Inlier as entirely ensialic with two or three successive rift events. This is supported by the chemical signatures of dolerites of various ages in the inlier, which, although they are generally low in Ba, Rb, Sr and K2O, are comparable to modern-day continental tholeiites (Ellis & Wyborn, 1984).
Episodic extensional tectonics have affected the central and eastern parts of the Mount Isa Inlier (Table 2.1). The first recognizable rifting event to affect the inlier (D$_{el}$) resulted in the deposition of coarse clastic sediments and felsic volcanics of the Argylla Formation around 1780 Ma. (Blake et al., 1985; Blake, 1986, 1987), and, as argued in this chapter, possibly the lower SCG. Ongoing crustal extension led to a change from felsic volcanism to mafic upper crustal igneous processes during the deposition of the Marraba Volcanics and Mitakoodi Quartzite, and upper SCG (Fig. 2.10).

As McKenzie (1978) and Beaumont (1978) have shown theoretically, during the period of tectonic quiescence and thermal relaxation that follows continental extension, subsidence occurs and low-energy "sag" sediments are deposited. These are laterally very continuous and lack the extensive extrusive intercalations and syn-sedimentary faults of the rift phase. A blanket of low-energy shelf sediments, including those of the upper Malbon Group, the Overhang Jaspilite, the lower Marimo Slate and maybe the lower Corella Formation has been interpreted as a sag sequence (Blake et al., 1985; Blake, 1986, 1987). This sequence generally lacks extrusives.

The Soldiers Cap Group (and its southwestern correlative the Kuridala Formation) contains the only major body of meta-turbidites in the Mount Isa Inlier. Since the SCG also delineates the eastern margin of the inlier, it has often been argued that this eastern margin represents a Middle Proterozoic continental margin (Plumb & Derrick, 1975; Wilson, 1978; Plumb et al., 1980). Blake et al. (1984) disputed this, arguing in favour of a source rock for the southern SCG towards the east, because, at least in the southern SCG, both the grain size and the feldspar content decrease towards the west (a full review of the pros and cons of the "continental margin" problem can be found in Blake, 1986, and a brief review in chapter 1). To this, one may add and emphasize that turbidites are not restricted to continental slope environments, although the continuity of the turbidites in the SCG seems to preclude deposition in a small and/or bathymetrically irregular basin. The base of the SCG is exposed south of the studied area: Laing & Beardsmore (1986) and Beardsmore et al. (1988, in press) have documented two lower units, which conformably underlie the SCG. A general progressively upward maturing of the meta-sediments and a marked upward change from bimodal, subaerial, felsic volcanism to subaqueous, mafic volcanism are the typical hallmarks of a rift-and-sag sequence. The SCG contains no MORB-type basalts. The depositional age of the SCG is poorly constrained but it is most likely (in all published correlation schemes) older than the meta-sediments in the Marimo-Staveley Block. Thus, there is no basis for a palinspastic reconstruction, which incorporates an eastward progressive deepening. More westwards, the Marimo-Staveley Block lithologies overlie those of the Mitakoodi Anticlinorium, so here too eastward deepening cannot be deduced. This adds to other
Tectonic evolution

A new and spatially restricted phase of magmatic activity was superposed on, and disturbed, the first sag phase: the Corella Formation northeast of the Mitakoodi Anticlinorium is intruded by the Burstall Granite of the Wonga Batholith (1745±17 Ma., Page, 1983b), whereas in the Tommy Creek Block, meta-sediments of the Corella Formation interfinger with ignimbritic sheets (Hill, 1987). It seems likely that this magmatic activity is due to a second pulse of extension. This second extensional event (De2) must have postdated the deposition of the Corella Formation, which it affects. The absolute maximum age of the Corella Formation is 1780 Ma., the U-Pb Zr age of the underlying Argylla Formation (Page, 1978, 1983a). However, as the Burstall Granite intrudes a large section of the Corella Formation, and sag sequences like the Corella Formation are deposited in slowly subsiding basins (time-constant for thermal relaxation of an extended lithosphere is in the order of 60 Ma.; McKenzie, 1978; Jarvis & McKenzie, 1980), it can reasonably be assumed that De2 did not start before 1760 Ma.. De2 must have predated ~1740 Ma., the recently determined age of a granite in the zone of extensional structures in the Mary Kathleen Fold Belt which itself is not affected by these extensional structures (R.W. Page, pers. comm., 1988). Thus, the age of this second extensional event must fall between 1760 and 1740 Ma., and it interfered with the thermal relaxation of the first, 1780 Ma. old, extensional event. Both purely orthogonal extension and purely orthogonal strike-slip-dominated tectonics (or a hybrid form) at De2 can explain the irregular bathymetric patterns that controlled the deposition of the upper Marimo Slate, Tommy Creek Block lithologies and Roxmere Quartzite. Dolerites intruded during this event have, like all other dolerites in the inlier, a continental tholeiitic signature. There is no evidence for the formation of oceanic crust at any stage during De2.

A third extensional event (De3) affected the western part of the -now exposed- inlier, and resulted in the deposition of the 1678±1 Ma. old Carters Bore Rhyolite (Page, 1978), the Surprise Creek Formation, and the 1670 Ma. old Mount Isa Group (Fig. 1.3; Page, 1981). Again, there is no evidence for the formation of oceanic crust. Thus, an (early, basin-forming) ensialic history for the inlier is very strongly supported.

De1 is a local thrust event, recognized in the Marimo-Staveley Block and the Soldiers Cap Group (in the Mitakoodi Anticlinorium it is locally manifested by a slaty cleavage) and between the Mitakoodi Anticlinorium and the Tommy Creek Block. Its movement vector pattern is erratic (compare Bell, 1983; BMR, 1985; Loosveld & Schreurs, Appendix 1; Loosveld, Chapter 3). Tear faults which could geometrically account for the different movement vectors have as yet not been recognized. De2 is the
penetrative deformation event in the central eastern Mount Isa Inlier (as in the rest of the inlier). $D_{c2}$ gave rise to the most penetrative map-scale structures. These structures were formed by coaxial E-W compression and vertical extension, resulting in a plane-strain shortening of 35-55%. $D_{c3}$ was a phase of N-S to NE-SW directed compression, leading to local refolding and to juxtaposition of tectono-stratigraphic blocks of different metamorphic grade. The youngest structures are kinks and faults.

The eastern part of the Mount Isa Inlier is a poly-metamorphic terrane. Peak-metamorphism is syn-$D_{c1}$ to pre-$D_{c2}$. Crustal thickening during $D_{c2}$ was, however, not accompanied by high-P facies metamorphism and andesitic volcanism, but by low-P facies metamorphism. It thus appears that the basin-modifying tectonics of the inlier, $D_{c1}$, $D_{c2}$, and $D_{c3}$, were also characterized by intra-continental processes, rather than by subduction-related processes. In chapters 3 and 4, these compressional events are further described and discussed.

A problem which needs to be addressed by geodynamicists now is the one of the compressional driving force, in the absence of preceding formation of oceanic crust. In the plate tectonics theory, extension and the formation of oceanic crust is logically followed by cooling of the oceanic lithosphere and subsequent destruction of this lithosphere. Continental collision may result. No such oceanic crust appears to have formed in the Proterozoic terranes of Australia (review by Etheridge et al., 1987b). Moreover, in the following chapters (Chapter 3 & 4), the timegap between the extensional events and the major compression ($D_{c2}$) will be shown to exceed 120 Ma., so that it seems unlikely that compression is causally linked to the extension at all.
CHAPTER 3

THE STRUCTURAL/METAMORPHIC FRAMEWORK OF THE CENTRAL SOLDIERS CAP GROUP BELT

LOOSVELD

Accepted by "Tectonophysics"; here modified for thesis
THE STRUCTURAL/METAMORPHIC FRAMEWORK OF THE CENTRAL SOLDIERS CAP GROUP BELT

3.1 INTRODUCTION

In this chapter, the relation between the structural and metamorphic histories of the central Soldiers Cap Group belt will be discussed. Specifically, the synchronicity of the growth of low-P facies metamorphic assemblages with $D_{c2}$, the penetrative, regional deformation event, will be demonstrated. In chapter 4, a general discussion of the possible tectono-magmatic models that may explain low-P facies metamorphism during a phase of crustal thickening will follow.

The structure of the western and central parts of the Mount Isa Inlier is generally characterized by upright, N-S trending folds, widely attributed to $D_{c2}$ (Derrick et al., 1977c; Bell, 1983; Lister et al., 1986b; Winsor, 1984, 1985, 1986; Loosveld & Schreurs, Appendix 1). In chapter 2 (and also by Donchak et al., 1983; Hill, 1987), it was argued that the characteristics of $D_{c2}$ in the three tectono-stratigraphic domains of the eastern part of the inlier, including the Soldiers Cap Group belt (Fig. 3.1), are similar to those in the central and western parts. Consensus on the pre-$D_{c2}$ history, however, is still far-off. Several different types of early structures have been described west of the Pilgrim Fault/Ballara-Corella River Fault Zone (Fig. 3.1). Passchier (1986a) interpreted extensional detachment-like geometries around the Alligator Syncline (Fig. 3.1); a penetrative (extensional?) fabric in the Wonga Granite also predates $D_{c2}$ (Pearson et al., 1987; Oliver et al., in prep.). Bell (1983) interpreted a huge thrust duplex with marginal synclines for the area north of Mount Isa township, and Loosveld & Schreurs (Appendix 1) attributed an imbricate stack of thrust sheets between the townships of Mount Isa and Mary Kathleen to $D_{c1}$. The temporal and spatial relations between these pre-$D_{c2}$ structures are not clearly established.

In the eastern part of the inlier, i.e. east of the Pilgrim Fault, the situation is markedly different, and even more obscure. Not only are pre-$D_{c2}$ structures poorly documented and correlated, the nature of $D_{c2}$ itself is also a matter of discussion. $D_{c2}$ is more varied in orientation and only very few pre-$D_{c2}$ structures have been documented and some of these have dubious status. Map-scale pre-$D_{c2}$ structures dominate an E-W strip, centred on the Cloncurry township. West of Cloncurry, where the zone is about 5 km wide, it comprises highly non-cylindrical folds and strong LS-fabrics (Hill, 1987). East of Cloncurry, a 10 km wide strip is characterized by tight to isoclinal, upright, E-W trending folds, which have been attributed to $D_{c1}$ (Glikson and Derrick, 1970). Ryburn et al. (in prep.) also describe $D_{c1}$ thrust planes and $D_{c1}$ folds east of Cloncurry, but attribute
the E-W folds to $D_{c2}$. Much of the confusion springs from the emphasis given to orientation while correlating structures.

Fig. 3.1 Location of the Soldiers Cap Group in the Mount Isa Inlier. Broad arrows represent $D_{c1}$ thrust movement vectors. Broken arrow represents the southwards component of the $D_{c1}$ movement vector (see §3.4, "$D_{c1}$ Movement Direction").
Recent work on thin-skinned thrust systems (among others Dahlstrom, 1969; Elliott, 1976a, 1976b; special issue Journ. of Str. Geol., Vol. 8, No. 3/4, 1986) has shown that one apparently simple compressive event can produce a wide range of structural styles with widely varying trends, and complexity is even greater if the more ductile processes at depth are considered. Careful mapping of cleavage traces provides a means of unravelling such complex geometries (Mitra & Elliott, 1980). This procedure was adopted during my 1986 and 1987 field seasons in the central Soldiers Cap Group belt, and the result demonstrates the simultaneous development of different fold trends that had previously been attributed to two separate fold phases. In brief, the structure, as it is interpreted in this chapter, is one of a refolded fold nappe: an early NW-closing fold nappe with an E-W trending, coaxially shortened frontal/lateral zone ($D_{c1}$), folded around a N-S trending, upright syncline ($D_{c2}$). Metamorphic grades in the study area are in places sufficiently high for large porphyroblasts to have grown, and micro-structural porphyroblast-foliation relations add to, clarify and confirm field relations. A 1:100 000 scale structural map of the central SCG belt is in appendix 9 (back pocket).

3.2 GENERAL SETTING

The Soldiers Cap Group (SCG) occupies the easternmost part of the Mount Isa Inlier. It is exposed in a NNW-trending belt, almost 250 km long and between 10 and 30 km wide. This structural study was carried out in the CLONCURRY 1:100 000 Sheet area, approximately in the middle of this belt, where the stratigraphic sequence is coherent and well documented (Carter et al., 1961; Glikson & Derrick, 1970; Glikson, 1972; Derrick et al., 1976c; Ryburn et al., in prep.). As already seen in chapter 2, the SCG can be subdivided into three formations, from base to top: the Llewellyn Creek Formation, the Mount Noma Quartzite and the Toole Creek Volcanics. The Llewellyn Creek Formation, of which the base is not exposed in the studied area, consists mainly of well-bedded meta-turbidites. The Mount Noma Quartzite grades upwards from reworked meta-turbidites and massive, clean meta-arenites into meta-siltstones interspersed with metabasalts. The Toole Creek Volcanics contain mainly metabasalts, intercalated with limestones, and fine-grained sediments such as meta-siltstones, phyllites and cherts. The total thickness of these three formations, as exposed in the Cloncurry 1:100 000 Sheet area, is about 6 km.

Also discussed in chapter 2 is the age of the SCG and its stratigraphic correlation with units to the west (Blake et al., 1983, 1984; Beardsmore et al., in press; Ryburn et al., in prep.). These are uncertain at present although a correlation with the Kuridala Formation seems justified (Blake et al., 1983; Laing & Beardsmore, 1986; Beardsmore et al., in press). The only direct geochronological control is given by the young, post-tectonic plutons of the approximately 1500 Ma. old Williams and Naraku Batholiths.
(Nisbet et al., 1983; R.W. Page, pers. comm., 1986), namely the Saxby Granite in the south and the Naraku Granite in the north (Fig. 3.2), which intrude the SCG. In chapter 2, on indirect and unfortunately non-conclusive evidence, I have argued for an age of approximately 1780-1750 Ma. for the SCG.

**Fig. 3.2** Map of the studied area. Isolated arrows designate the orientations of the $L_1$ extension lineations. "A" and "B" are localities of respectively Figs. 3.4A and 3.5. At locality "C", a vertical E-W trending crenulation cleavage is developed; at "D" a steeply SSW-dipping differentiated crenulation cleavage.
depending on: Al\textsuperscript{III} – Si\textsuperscript{III} disorders (Greenwood, 1972) (Graham & Williams 1985)

- solid solution of minor constituents
- enthalpy change of the orth. → sill. reaction (Anderson et al., 1977)
A detailed metamorphic study of the southern SCG belt has been undertaken by Jaques et al. (1982). Except for a slightly higher-grade first metamorphic event in the south (sillimanite/K-feldspar), their observations and conclusions can be extended northwards into the central SCG belt. The meta-turbidites of the Llewellyn Creek Formation and the Mount Noma Quartzite show signs of two distinct metamorphic events: the first one was a low to moderate-pressure/high-temperature, prograde regional event with diagnostic minerals such as garnets, andalusite, staurolite, chloritoid and sillimanite in pelitic rocks, and anthophyllite in the more arenitic parts of the meta-turbidites (Table 3.1). Just west of the axis of the Snake Creek Anticline and north of the Saxby Granite (at CLONCURRY G.R. 650791), up to 2 cm long sillimanite pseudomorphs (or partial pseudomorphs) after chiastolitic, tetragonally prismatic andalusite porphyroblasts are present (sample L200). The persistence of andalusite in this area (and also in the sillimanite-grade core of the Middle Creek Anticline) is probably due to its meta-stability (\textasciitilde 50°C). Anthophyllite + cordierite + garnet in metamorphosed dolomitic (?) marls is reported by Jaques et al. (1982). The co-existence of staurolite and andalusite constraints the metamorphic gradient to a minimum 41°C/km (using equilibrium conditions after Richardson, 1968, Hoschek, 1969, and Holdaway, 1971, and assuming a specific density of the crust of 2700kg/m³; Fig. 3.3). The maximum metamorphic temperature is defined by the intersection of the water-saturated granite melting curve (Merrill et al., 1970) and the "muscovite + quartz K-feldspar + sillimanite + H_2O" phase boundary (Helgeson et al., 1978) at 665°C. No detailed geothermometry and geobarometry studies have been undertaken in this Ph.D project, but the similarity of mineral assemblages (Table 3.1) in the central SCG belt with those of the southern SCG belt (Jaques et al., 1982) indicate similar P/T ratios for the prograde sequence. The shaded area in Fig. 3.3 represents the approximate peak-metamorphic pressure and temperature conditions of the SCG. Jaques et al. (1982) calculated metamorphic pressures of 3-4kbar and temperatures of 450-680°C for the prograde metamorphism. The second metamorphic phase was retrograde in character, its highest grade defined by biotite (\textasciitilde 6, 6, 6). Except for the sillimanite-in isograd in the Snake Creek Anticline area (Fig. 3.2 & Appendix 9), isograds have not been accurately mapped in this study. Still, it appears that isograds are approximately parallel to bedding. For example, in the Snake Creek Anticline area, sillimanite is only found in the Llewellyn Creek Formation, whereas andalusite-staurolite-garnet schists generally occur from the sillimanite isograd upwards for about one kilometre into the Mount Noma Quartzite. Biotite, muscovite and chlorite are diagnostic minerals for the upper Mount Noma Quartzite and the Toole Creek Volcanics. In the Middle Creek Anticline area (Fig. 3.2; Appendix 9), the isograd
Table 3.1 Mineral assemblages of the central SCG belt. Notation of phases: ab, albite; act, actinolite; and, andalusite; anth, anthophyllite; bi, biotite; cc, calcite; chl, chlorite; cpx, clinopyroxene; epi, epidote; fs, feldspar; ga, garnet; hbl, hornblende; ilm, ilmenite; mu, muscovite; pl, plagioclase (An, anorthite-percentage); qtz, quartz; sill, sillimanite; st, staurolite.

<table>
<thead>
<tr>
<th>Pelitic</th>
<th>Psammite</th>
<th>Basic</th>
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<tbody>
<tr>
<td>qtz + bi + mu + si + ga</td>
<td>qtz + fs + ga ± mu</td>
<td>pl (An 25-55) ± hbl ± cpx ± ga ± ilm</td>
</tr>
<tr>
<td>qtz + bi + mu + ga ± fs</td>
<td>qtz + fs ± bi ± mu</td>
<td>pl + hbl + epi + ilm + qtz ± act</td>
</tr>
<tr>
<td>qtz + bi + mu + st + ga</td>
<td>qtz + mu + bi ± ga ± fs</td>
<td>qtz + ab + epi + chl ± hbl</td>
</tr>
<tr>
<td>qtz + bi + mu ± fs</td>
<td>qtz + mu + bi ± fs ± anth ± cc</td>
<td>qtz + fs + mu + chl</td>
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<tr>
<td>qtz + bi + mu + st + and</td>
<td>qtz + fs + mu + chl</td>
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<td>qtz + bi + mu + ga + and</td>
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<td>qtz + bi + mu + and + fs</td>
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<td>qtz + mu + chl</td>
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Fig. 3.3 P-T diagram showing the approximate equilibrium conditions for the prograde metamorphism of the lower SCG in the central SCG belt. Reaction (1) lower staurolite limit (after Richardson, 1968); (2) staurolite + muscovite + quartz ↔ Al₂SiO₅ + biotite + H₂O (Hoschek, 1969); (3) muscovite + quartz ↔ K-feldspar + Al₂SiO₅ + H₂O (Helgeson et al., 1978); and (4) water-saturated granite melting curve (Merrill et al., 1970).
pattern is more complex, as sillimanite has not only been reported from the core, but also from isolated occurrences away from the core (S. Walters, pers. comm., 1987).

The approximate bedding-parallelism of isograds in the Snake Creek Anticline area might be interpreted to indicate that this prograde metamorphic event took place early in the deformational history. However, it can be unambiguously demonstrated that the prograde metamorphism is post-$D_{c1}$ and probably early-$D_{c2}$, since its porphyroblasts have overgrown the penetrative $D_{c1}$-fabric, which is crenulated to a lesser degree within the porphyroblasts than it is in the matrix (Fig. 3.4A). Non-equidimensional minerals like biotite and fibrolitic sillimanite mimic the (crenulated) $D_{c1}$-fabric. Sillimanite and biotite laths commonly crosscut each other in the $D_{c2}$ crenulations, instead of being passively folded around the $D_{c2}$ crenulations. Similar porphyroblast-foliation relations have been documented for other parts of the inlier (brief review in chapter 2). The $D_{c1}$ LS-fabric (Fig. 3.5) in both the pelitic and arenitic parts of the meta-turbidites of the SCG is microstructurally defined by flattened and elongated quartz crystals and aggregates, around which amphiboles (e.g. anthophyllite) and various micas anastomose. $L_1^1$, the extension lineation, in the Llewellyn Creek Formation and the upper Mount Noma Quartzite, consists of a variety of platy minerals (mica's, hornblende, ± fibrolitic sillimanite) which anastomose around elongate aggregates of quartz.

The absence in the sillimanite zone (and more southerly in the sillimanite/K-feldspar zone) of signs of nearby anatexis - which, considering the steep metamorphic gradients, should have occurred below these metamorphic zones - can potentially be explained by the Coulomb-Navier fracture theory: tensile (opening) fractures through which melt can migrate, cannot be vertical in a compressional regime. With overall high differential stress levels, melt migration will be severely constrained in all directions; if melt migrates at all, it will do so horizontally.

All post-$D_{c2}$ deformations postdate this prograde metamorphic event. In the higher-grade rocks, retrograde biotite, sericitic muscovite and chlorite are associated with younger (differentiated) crenulation cleavages. Quartz is recrystallized in the hinges of these crenulations. Polygonal networks of recrystallized quartz also overprint the $D_{c1}$-fabric. Porphyroblasts are commonly rimmed by replacement minerals, and sometimes completely replaced (e.g. garnet by biotite/chlorite; andalusite by secondary muscovite/quartz).
Fig. 3.4 A: Garnet blastesis postdates $S_1$ and is synchronous with the early
development of the $D_{c2}$ crenulation. Locality A on Fig. 3.2 (plane-polarized light).
B: Similar porphyroblast-foliation relations from the Tommy Creek Block (MARRABA
G.R. 150036; BMR thin section 86531729), where Hill (1987; in prep.) argued for a pre-
to syn-$D_{c1}$ metamorphism. BMR thin section 86531729 (crossed nicols).
For the remainder of this thesis, only two metamorphic factors are of importance. First, the peak-metamorphic event is of low-P facies character, which is obvious from the mineral assemblages (Fig. 3.3). Second, this metamorphic event is prograde (+dT/dt), with higher metamorphic grades during $D_{c2}$ than during $D_{c1}$, as peak-metamorphic assemblages overgrew $D_{c1}$ fabric elements and are synchronous with $D_{c2}$ (Fig. 3.4).
3.3 STRUCTURE

In the Mount Isa Inlier, N-S trending folds are generally regarded as Dc2 structures (Bell, 1983; Lister et al., 1986b; Winsor, 1986; Etheridge et al., 1987b; Loosveld & Schreurs, Appendix 1). In the central SCG belt, however, strictly N-S trending folds are relatively rare (Fig. 3.2): only the Weatherly Creek Syncline and few km-scale folds which overprint the E-W trending Toole Creek and Oonoomurra Synclines (Appendix 9) trend N-S. The Snake Creek Anticline trends NNW and the Middle Creek Anticline NE. The Pumpkin Gully Syncline trends largely E-W, but is refolded about a WNW-trending axis.

The Snake Creek and Middle Creek Anticlines received most attention in this project, as they contained the lower unit of the SCG, the Llewellyn Creek Formation, which has the highest metamorphic grade, and thus provides the best micro-structural control through porphyroblast-foliation relations. In the Snake Creek Anticline area, the most pervasive mesoscopic structure is a steeply ENE-dipping slaty cleavage, associated with isoclinal folds on all scales. This is also the oldest recognizable foliation, S1. A steeply SSE-plunging extension lineation, L1, is well developed on S1 (Fig. 3.6). The S1 and L1 fabric elements are not folded by the Snake Creek Anticline, but instead are axial planar to it. Therefore, the Snake Creek Anticline must be a Dc1 fold and not a younger "cross-fold".

This conclusion is supported by two factors of correlation. First, both S1 and L1 are overprinted by peak-metamorphic, low-P facies mineral assemblages. In other parts of the inlier, similar assemblages are considered to be syn-Dc2 (e.g. Reinhardt & Hamilton, in press). Second, the style of the Snake Creek Anticline differs markedly from the general style of Dc2 folds. Dc2 folds are upright and symmetric on km-scale, i.e. without large sheared out limbs. In contrast, the Snake Creek Anticline is reclined with both its western limb and hinge zone overturned, and asymmetrically deformed, its western limb being more strained than its eastern limb. In the eastern limb, S1 and bedding are generally at a distinct angle to each other, even in the pelitic tops of the meta-turbidites (refraction within a complete Ta-« meta-turbidite can amount to 30°). In the western limb, however, S1 is stronger developed, with bedding often boudined and transposed subparallel to S1. Rare asymmetric folds in the Llewellyn Creek Formation in the western limb show the same NW-vergence as on the eastern limb (see §3.4, "Dc1 Movement Direction"). On the western limb, parallelism of bedding with S1 and strongly non-cylindrical folds indicate high shear strains. Towards the west, this zone of high, ductile shear strain becomes gradually obscured by the so-called Cloncurry "Overthrust" (Honman et al., 1939; Carter et al., 1961), a young high-angle reverse fault system,
Fig. 3.6 Equi-angular, lower hemisphere projection of (A) the extension lineation, \( L_1 \), (B) of \( S_2 \) and \( S_3 \), and (C) of \( S_1 \). The three great-circles in each stereoplot represent the "mean" axial planes of respectively the Middle Creek Anticline (long dashes), the Weatherly Creek Syncline (short dashes) and the Snake Creek Anticline (extra long dashes plus dots), calculated from the Bingham axial distribution of respectively \( S_2 \) in the Middle Creek Anticline, \( S_2 \) in the Weatherly Creek Syncline, and \( S_1 \) in the Snake Creek Anticline.
which separates the andalusite-staurolite schists of the SCG in the east from the
greenschist-grade Marimo-Staveley Block in the west (Fig. 3.2). This fault system
postdates intrusion of the main phase of the Williams Batholith (1509±22 Ma., Nisbet
et al., 1983).

$D_{c2}$ and $D_{c3}$ in the Snake Creek Anticline are principally represented by open to tight
crenulations of $S_1$ in the meta-pelites and irregularly spaced cleavages in the meta-arenites.
$D_{c2}$ structures trend N to NNW, whereas $D_{c3}$ crenulations trend NW to WNW (Fig. 3.6).
In some cases, short and narrow, evenly spaced shear planes transect $S_1$, such that an
extensional crenulation cleavage (Platt & Vissers, 1980) results. A km-scale $D_{c3}$ fold is
developed just north of the Saxby Granite (KURIDALA G.R. 672777, Appendix 9).
Pegmatite dykes, 2 to 30 cm thick, are folded around its moderately to steeply SSW-
dipping axial planes, thus indicating a post-intrusion age for $D_{c3}$. In the axial-plane zone,
strong $D_{c3}$ crenulations are developed. In the northernmost Snake Creek Anticline area
and the Toole Creek Syncline area, another system of probably post-metamorphic
crenulations consistently dips moderately to steeply towards the northwest. Correlation
of these with other structures failed, as they are only developed in bands and cannot be
traced laterally.

$S_1$, the pervasive slaty cleavage, can be continuously traced, in a clockwise rotation,
from the core of the NNW-trending Snake Creek Anticline to the Toole Creek
Syncline and its adjacent E-W trending folds, where $S_1$ remains the axial plane to the E-
W folds (also in the Oonoomurra Syncline; Fig. 3.7A; Appendix 9). Thus, the Toole
Creek Syncline is a $D_{c1}$ fold too. Slaty cleavages which were reported to be folded
around the E-W trending folds (R. Ryburn, pers. comm., 1986; B. Skrzeczynski, pers.
comm., 1986) appear to be in disagreement with this conclusion. Close examination of
the hinges in the Toole Creek fold system, however, revealed that the "folded" slaty
cleavage is in fact in most cases a slaty axial-plane cleavage to these hinges. Extreme
convergence of slaty cleavages in the incompetent slates gives a false impression of a
folded cleavage; slaty cleavage and bedding are (sub-)parallel except in the very hinges,
which are often abrupt. A vertical, E-W trending axial-plane crenulation cleavage
overprinting a slaty cleavage at CLONCURRY G.R. 635033 (point "C" in Fig. 3.2; inset
"B" at appendix 9) also seemed to imply a post-$D_{c1}$ origin for the E-W trending folds (R.
Ryburn, pers. comm., 1986; B. Skrzeczynski, pers. comm., 1986). However, only one
other fold hinge with a crenulation cleavage in the axial plane was found in the generally
E-W trending fold system (at CLONCURRY G.R. 665063; point "D" in Fig. 3.2;
Appendix 9). These two anomalous fold hinges are in this thesis interpreted as resulting
from progressive refolding during $D_{c1}$ ($S_{1c}$ over $S_{1A}$ in the terminology of Loosveld &
Schreurs, Appendix 1).
The Weatherly Creek Syncline (Fig. 3.2; Appendix 9), east of the Snake Creek Anticline, is the largest N-S trending fold in the central SCG belt. It plunges gently to steeply towards the north. Its parasitic folds are highly non-cylindrical. In the Cloncurry 1:100 000 Sheet area, it only exposes the Toole Creek Volcanics. Since the slaty cleavage, $S_1$, is not homogeneously developed in this topmost SCG formation, few crenulations exist in this syncline. Fold/foliation overprinting patterns, however, do occur; a weak slaty cleavage and accompanying dm-scale, N-plunging, tight fold obliquely transect isoclinal, steeply SW-plunging folds in the eastern limb of the Weatherly Creek Syncline (CLONCURRY G.R. 744878). Type II fold overprinting patterns ("mushroom", Ramsay, 1967) result (sample L48). Also, $L_1^1$ is inferred to be folded around the Weatherly Creek Syncline: although the actual folding of the extension lineation cannot be observed in the low-grade Toole Creek Volcanics in the hinge of the syncline, since the lineation is not clearly developed here, to the west, in the Snake Creek Anticline, it consistently plunges towards the S or SE, whereas to the east, in the western limb of the Middle Creek Anticline (Figs. 3.2 & 3.6A; Appendix 9), it plunges towards the north. South of the Saxby Granite, where higher-grade metamorphic meta-turbidites of the Mount Noma Quartzite and Llewellyn Creek Formation are exposed in the extension of the Weatherly Creek Syncline, $S_1$ is (sub-) horizontal in $D_{c2}$ hinges with $L_1^1$ trending NW on it. The axial-plane slaty cleavage is overprinted by NW-trending crenulations. In conclusion, the Weatherly Creek Syncline must be a post-$D_{c1}$ structure, and, by comparison with tight to isoclinal, N-S trending, and non-cylindrical $D_{c2}$ folds in other parts of the inlier, is most likely a $D_{c2}$ fold.

The Middle Creek Anticline (Fig. 3.2; Appendix 9) is the easternmost structure exposed in the Mount Isa Inlier. Exposure rapidly decreases eastwards. The Broken Hill Proprietary Co. Ltd. provided (1) pre-processed aero-magnetic and aero-radiometric imageprints, (2) the raw aero-magnetic and aero-radiometric data sets and their image-processing laboratory in Camberwell, Melbourne, and (3) their diamond drill core, thus greatly facilitating mapping in the poorly to non-exposed eastern limb (Fig. 3.7B). After interpreting a wide range of mainly "aero-magnetic" imageprints (many of them obtained by sun-illumination image processing), the Middle Creek Anticline was, in the 1987 field season, mapped on 1:25 000 scale aerial photographs. It is a moderately NE-plunging, moderately to steeply NW-inclined, tight to isoclinal anticline, with Llewellyn Creek Formation in the core and Mount Noma Quartzite around it. A penetrative, NE-dipping (dip direction 315±45) crenulation cleavage is developed all over the anticline (Fig. 2.9B). Slaty cleavages are progressively stronger developed towards the core of the anticline, and are in places axial-planar to isoclinal folds, with a vergece which does not fit the Middle Creek Anticline. In the Middle Creek Anticline (Fig. 2.1), only the upper limb of the Snake Creek fold nappe re-emerges. A "christmas tree" fold overprinting
Deformation/metamorphism pattern (type II, Ramsay, 1967) emerged, with SW-verging Dc1 folds on the western limb of the Middle Creek Anticline and NW-verging Dc1 folds on the eastern limb, and with the crenulation cleavage as axial plane of the Middle Creek Anticline. Appendix 9 shows the relation between bedding, slaty cleavage and crenulation cleavage. Microscopic examination of the bedding-slaty cleavage-crenulation cleavage-porphyroblast relations supports the macro- and mesoscopic structural pattern, and once again confirmed that the low-P facies mineral assemblages are syn-Dc2. Two major WNW-trending, sinistral faults displace the hinge of the Middle Creek Anticline and are thus post-Dc2. Various kinks also affect Dc2 crenulations.

The Pumpkin Gully Syncline lies south(-east) of the Naraku Batholith (Fig. 3.2; Appendix 9). Its core is made up of metabasalts and minor siltstones of the lower Toole Creek Volcanics, its rim by meta-basalts, fine-grained quartzites, meta-siltstones and BIFs of the Mount Noma Quartzite and dolerites. It is surrounded in the north, west and south by calcisilicate breccias; in the east, the syncline disappears under a thin cover of Cenozoic sediments. Ryburn et al. (in prep., on their final CLONCURRY 1:100 000 Sheet) interpreted the calcisilicate breccias, which they named the Gilded Rose Breccia, as being derived from the Corella Formation. The complete syncline overlies the calcisilicate rocks, and, as the SCG is older than these, a thrust is inferred between the calcisilicate breccias and the SCG. (On the same ground, a thrust has been inferred at the base of the Mount Noma Quartzite in the syncline immediately east of Cloncurry township). The hinge zone of the Pumpkin Gully Syncline around CLONCURRY G.R. 648116 is particularly instructive with respect to the series of structural events (inset "A" in appendix 9). Isoclinal rootless folds dominate the BIF in the hinge of the Pumpkin Gully Syncline; a strongly developed, bedding-parallel slaty cleavage is developed in the adjacent siltstones of the Mount Noma Quartzite and is folded around the hinge of the syncline. The rootless, isoclinal folds and the slaty cleavage are the oldest recognizable fabric elements in the syncline (Dc1). They are overprinted by vertical NW-SE trending crenulation cleavages, interpreted here as Dc2 structures. Tracing the crenulation cleavage along the hinge into the Toole Creek Syncline, it gradually is replaced by a slaty cleavage.

Fig. 3.7 A: Aerial photograph RC 9 (CAB 301) run 71150, showing the traces of S1 as developed in the Toole Creek Syncline fold system and the northern Snake Creek Anticline (Crown Copyright; reproduced by permission of the Director, Division of National Mapping, Department of Resources and Energy, Canberra, Australia). B: Imageprint of aero-magnetics from the Snake Creek Anticline, the Weatherly Creek Syncline, the Middle Creek Anticline and the Toole Creek Syncline fold system (produced with assistance of Wayne Stasinowski; reproduced by permission of "The Broken Hill Proprietary Co. Ltd.").
implying that $D_{c1}$ was restricted to the rim of the syncline. This may well fit the inferred thrust at the contact between he SCG and the calcisilicate rocks. The entire syncline is folded around a WNW-trending fold, which must be post-$D_{c2}$, and which is not demonstrably associated with any mesoscopic structures. A sinistral, NNW-trending fault is also post-$D_{c2}$, but the relation to the young fold is uncertain as the fault only affects the northern limb of the WNW-trending post-$D_{c2}$ fold. Although none of the fabric elements can be traced over the calcisilicate breccias to the southern structures, $D_{c1}$, $D_{c2}$ and post-$D_{c2}$ structures in the syncline show sufficient similarities with respectively $D_{c1}$, $D_{c2}$ and $D_{c3}$ structures in the southern part of the central SCG belt area to be positively correlated with them.

Three structural correlation schemes have been proposed. Glikson & Derrick (1970) regarded the E-W trending, vertical to steeply S-dipping, tight to isoclinal Toole Creek Syncline fold system as a $D_{c1}$ fold system connecting southeastwards with the N-S trending Weatherly Creek Syncline. This $D_{c1}$ Weatherly Creek/Toole Creek Syncline was, in their view, folded by a set of "cross-folds", of which the Snake Creek Anticline was the largest. Recently, R. Ryburn (written comm., 1986) put forward an alternative deformation history: again the Weatherly Creek and the Toole Creek Synclines were regarded as part of the same fold generation, $D_{c2}$ this time, postdating a $D_{c1}$ Snake Creek Anticline and postdated by a NE-trending $D_{c3}$ fold. This was based on observations made around various W-closing, E-plunging, tight synclines in the Toole Creek fold system, where an early foliation is folded around these hinges. Since then, Ryburn et al. (in prep.) have prepared the final CLONCURRY 1:100 000 Sheet; on it, they show the Toole Creek Syncline as a $D_{c1}$ fold, folded around a $D_{c2}$ Snake Creek Anticline, whereas the Weatherly Creek Syncline is also interpreted as a $D_{c2}$ fold.

Briefly, the following points, obtained from the structural mapping (Appendix 9), argue against these three schemes:

1. $S_1$ and $L_1$ are not folded by the Snake Creek Anticline, but are axial planar to it. They are overprinted by syn-$D_{c2}$, peak-metamorphic, low-P facies mineral assemblages. The Snake Creek Anticline is a $D_{c1}$ synformal anticline and not a younger "cross-fold".
2. $S_1$ can be continuously traced from the Snake Creek Anticline to the Toole Creek Syncline fold system. The Toole Creek Syncline is also a $D_{c1}$ fold as $S_1$ is the axial plane to it.
3. The Weatherly Creek Syncline is a post-$D_{c1}$ structure, because $S_1$ and $L_1$ are folded around it. By inference, it is here interpreted as a $D_{c2}$ fold.

$D_{c1}$ is characterized by completely different structures in two domains. In the Snake Creek domain one large, NWW-trending, synformal anticline crops out, probably rotated...
into that orientation during $D_{c2}$. Originally, it may have been a NNW-closing recumbent or inclined fold. In the Toole Creek Syncline domain, on the other hand, a system of tight to isoclinal, E-W trending folds occur, which are only gently affected by $D_{c2}$ (Figs. 3.2 & 3.7A). Still, $S_1$ can be traced from the originally recumbent or inclined Snake Creek Anticline northwards into the upright E-W trending Toole Creek fold system. Figure 3.8A illustrates the pre-$D_{c2}$ geometry. Thus, a simple shear dominated, recumbent domain (Snake Creek) passes into a domain which is dominated by coaxial horizontal N-S shortening (Toole Creek) and the picture of a fold nappe with a coaxially shortened front emerges. This concept is not a new one. Heim (1919) observed that upright folds commonly occur at the front of a fold nappe. Here, the geometry may be more complicated, as the $D_{c1}$ movement vectors are not exactly perpendicular to the Toole Creek Syncline fold system ($§3.4$). The Toole Creek Syncline fold system may be an obliquely shortened zone.

If "reflection" symmetry exists along the axial plane of the Weatherly Creek Syncline, the area around the Middle Creek Anticline (Fig. 3.2) should provide another section through the $D_{c1}$ fold nappe. However, the lower limb of the $D_{c1}$ structure is not exposed here. SW- to NW-verging $D_{c1}$ folds are all considered to represent the same $D_{c1}$ (upper) limb, and are consistently overprinted by moderately to steeply NW-dipping $D_{c2}$ crenulation cleavages, which parallel the axial plane of the Middle Creek Anticline (Fig. 3.6). However, if the $D_{c1}$ fold nappe extended laterally (approximately north-eastwards) for some considerable distance during $D_{c1}$, the décollement might reappear somewhere to the east of the Middle Creek Anticline (Fig. 3.2). This area, however, has extremely poor exposure, inhibiting verification.

In the model proposed here, it is suggested that the post-$D_{c2}$ Cloncurry "Overthrust" locally reactivated the folded $D_{c1}$ décollement-type shear zone, which underlies this Snake Creek fold nappe. Structural analyses that neglect the tracing of foliations, particularly $S_1$, but in which structures are correlated by orientation, lead to completely different, and in my view erroneous, results.

### 3.4 $D_{c1}$ MOVEMENT DIRECTION

Determination of the $D_{c1}$ movement vector in the central SCG belt is difficult, because of the overprinting of $D_{c1}$ sense-(or direction-)of-shear indicators by later structural/metamorphic fabric elements and rotation during later folding. As the mechanisms of later folding ($D_{c2}, D_{c3}$) are at present a matter of debate, backrotation of $D_{c1}$ structures to their original position is at best speculative. The variable rotation of $S_1$ with respect to $S_0$ during $D_{c2}$ and $D_{c3}$, and the rotation of $L_0^1$ towards $L_1^1$ and $L_2^1$ during

$L_x^{S_1}$: foliation intersection lineation

$L_x^*$: extension lineation
Fig. 3.8 A: Schematic sketch of the $D_{1}$ fold nappe with a "crumpled up" zone in front of it. The recumbent fold in the foreground represents the Snake Creek Anticline, the upright fold system the Toole Creek Syncline. B: Schematic representation of the present configuration.
respectively $D_{c1}$ and $D_{c2}$ further hamper backrotation of $D_{c1}$ fabric elements. Also, in pelitic (incompetent) units of the lower SCG, the commonly small angle between $S_0$ and $S_1$ hinders recognition of the $L_0$ intersection lineation, and in places the optical illusion of vergence reversal, described by Williams (1985), obscures the sense of vergence. Finally, in these pelitic units, the anastomosing of $S_1$ around and through large porphyroblasts results in a chaotic crenulation pattern.

In $D_{c2}$ hinges, where, assuming plane strain during $D_{c2}$, the orientation of $L_1$ is unaltered by $D_{c2}$, $L_1$ is orientated in a NW direction, indicating a NW-SE direction of shear (e.g. KURIDALA G.R. 802662). Apart from some extensional crenulation cleavages of doubtful origin, there are no S-C relations, which might give the $D_{c1}$ sense of shear. The approximate sense of shear in the Snake Creek Anticline-Weatherly Creek Syncline area can be determined in three, unfortunately inaccurate and disputable, ways. Firstly, since the Snake Creek fold nappe closes in the north-northwest, and the $D_{c1}$ fold axes here are at a large angle to the $L_1$ extension lineations, one may assume that the sense of shear was towards the NNW. Secondly, by analogy to thrust nappe tectonics, where a décollement or thrustplane climbs through the stratigraphy in the direction of movement, it can be argued that, since the $D_{c1}$ high-shear strain zone immediately east of the Cloncurry "Overthrust" ($D_{c3}$) climbs from the Llewellyn Creek Formation in the south-southeast through the Mount Norna Quartzite to the Toole Creek Volcanics in the north-northwest, there is a component of the translation vector towards the north-northwest. Finally, the consistently NNW-verging parasitic $D_{c1}$ folds in both the upper and lower limb of the Snake Creek Anticline, may point to a NNW sense of movement (Fig. 3.8B).

The NNW-vergence of parasitic $D_{c1}$ folds on both limbs of the Snake Creek Anticline could be wrongly interpreted as being unrelated to the development of the Snake Creek Anticline or associated to more than one deformation event. There are, however, three possible explanations for the development of similarly verging $D_{c1}$ parasitic folds on both limbs of a major $D_{c1}$ fold. (1) In very strongly sheared rocks, bedding can be rotated (almost) into the shear plane, after which a perturbation in the flow field can result in unusually vergent folds. (2) Consistently NNW-verging folds can also be explained by simple shear as the mechanism controlling the $D_{c1}$ fold development on both the upper and lower limb as the nappe develops (Fig. 3.8A). For this to be the case, the bedding-anisotropy in the lower limb has to control (or be reactivated by) a local simple shear system, slightly deviating from the overall simple shear, with shear along the bedding planes, so that every irregularity on its surfaces can potentially develop into an asymmetric and non-cylindrical fold (Fig. 3.9). Bedding on the lower limb may then progressively approach the $XY$-plane of the overall strain ellipsoid, but it also coincides with the
dominant shear plane of the local incremental strain ellipsoids. (3) Additionally, relatively early Dc1 parasitic folds, that "roll" passively from the upper limb through the fold nappe hinge, will show roughly the same vergence on the lower limb. Since, however, no westward verging folds were observed in the hinge of the Snake Creek Anticline, this additional rolling hinge effect has probably not greatly influenced the parasitic fold geometry.

Fig. 3.9 Development of asymmetric parasitic folds of equal vergence in both the upper and the lower limbs of a fold nappe. A: late stage in the development of a fold nappe. B: the associated incremental strain ellipsoid. C: Enlargement from the short limb of the fold in "A". See text for explanation.

Determination of the Dc1 movement vector in the Middle Creek Anticline area yields a slightly more westwards orientation. Here, only the upper Dc1 limb is exposed, folded around the NE-plunging Dc2 Middle Creek Anticline. The Dc1-vergence is variable, but, after backrotation of Dc2, has everywhere a component towards the west. The backrotation applied here consists of two steps: first a passive rotation of L₀ through a vertical plane back to horizontal; L₀ and L₁ are rotated along the same horizontal rotation axis. Then, L₀ and L₁ are rotated along the new L₀ rotation axis until they are horizontal.
This way of backrotation is probably valid as the two clusters of $L_1$ (Fig. 3.6) converge to only one remaining cluster. This $L_1$ cluster indicates an ESE-WNW to E-W direction of shear. This combined with the generally westwards $D_{c1}$ vergences, gives a W(NW) sense of shear (the $L_0$ cluster remains, as expected with the non-cylindrical $D_{c1}$ folds, very wide after the backrotation).

3.4.1 Discussion of the regional $D_{c1}$ movement vector field

A major problem lies in the diversity of the reported $D_{c1}$ movement vectors in the Mount Isa Inlier (BMR, 1985; Fig. 3.1). Bell (1983) interpreted a huge duplex north of Mount Isa township with a southward sense of movement and more than 200 km of displacement. Loosveld & Schreurs (Appendix 1) confirmed this southwards sense of movement in an area northeast of Mount Isa township, but interpreted an east over west to southeast over northwest translation further east, some 15 km northwest of Mary Kathleen township. Hill (1987) argued for a SSE-directed $D_{c1}$ movement vector for the Tommy Creek area (Fig. 3.1). This study, in the central SCG belt, has yielded a W to NNW sense of shear. An as yet unsolved question is: are these different movement vectors related to different pre-$D_{c2}$ events, that should not be correlated? Or can a movement vector field of one thrust event be so capricious and exist of different domains?

The former possibility assumes several phases of horizontal compression, all affecting only one domain. One may question the feasibility of this, as it is unlikely that shortening is concentrated in small (non-linear) domains without affecting the surrounding domains (I am unaware of any publication documenting such a heterogeneous deformation). It is more likely that a capricious vector field is associated with one thrust event ($D_{c1}$), resulting in domains with different thrust directions. This is even more likely as at the onset of $D_{c1}$, the inlier had already suffered a complex, multiphase extensional history (including possible strike-slip deformation), and was presumably compartmentalised by numerous normal, transfer and detachment faults. These pre-existing faults, and additional lithological anisotropies, may have acted as zones of weakness, and may locally have diverted the overall thrust direction. Major geometrically necessary lateral ramps (Boyer and Elliott, 1982) or tear faults, analogous to extension-related "transfer faults" (Gibbs, 1984; Etheridge, 1987) or "accommodation zones" (Bosworth, 1985), must bound these domains. And like normal faults which can have different orientations across a transfer fault (Lister et al., 1986a), thrusts can be envisaged to change dip, and therefore have different movement vectors, across one or more tear faults. As yet only the Wonga Belt (Fig. 3.1) has been suggested to be such a tear fault (T.H. Bell, pers. comm., 1985; R.L. Hammond, pers. comm., 1985). Remarkably, it has also been explained in terms of a major $D_{c2}$ extensional zone (Pearson
Note: clonal metamorphism is unlikely on such a small scale.

Thompson's (1976) clonal metamorphism in the Simplon area (Switzerland) is not believed by Niggli (1970) and Milnes (1976), who argue for tectonic doming in the area.
et al., 1987; Holcombe et al., 1987). Other tear faults must be hidden in the Mount Isa Inlier. If movement vectors along both sides of such a tear fault are directly opposed, the tear fault must be (and must have been) vertical. Variations from 180° can be explained by non-vertical tear faults.

3.5 DISCUSSION AND CONCLUSIONS

Dc1 in the central SCG belt is manifested by widely varying structures. In this chapter, I have argued that the approximately NNW-trending Snake Creek Anticline (Fig. 3.2) is a Dc1 structure: S1 is the axial plane to it, and the mineral lineation on S1, L1, consistently plunges to the S or SE (Fig. 3.6). The Snake Creek Anticline differs from other N-S trending structures, which are generally upward-facing, in that most bedding planes in the hinge and in the western limb are slightly overturned, making it sensu stricto a steep synformal anticline. It is interpreted here as a refolded frontal fold nappe. The fold nappe can be traced through a transitional zone into a system of tight, upright folds in its frontal/lateral zone. Whereas the fold nappe domain was dominated by W- to N-directed progressive simple shear, the upright folds may be the result of coaxial N-S shortening.

A feature of this model which needs further discussion (and warrants further investigations) is the isograd pattern. Although isograds have not been accurately mapped, in the Snake Creek Anticline area, they are sub-parallel to bedding folded by Dc1. The sillimanite isograd e.g. lies just west of the axial plane of the Snake Creek fold nappe, but mimics this Dc1 fold nappe (Fig. 3.2; Appendix 9). Small deviations from perfect parallelism with Dc1 structures, like this "shifted" sillimanite isograd may reflect the shape of the isotherms during, or just after, the fold nappe formation during the sillimanite blastesis. When these are not yet completely readjusted to horizontal planes. In other words, the folded sillimanite "isotherm" had sunk below the now exposed axial zone of the Dc1 Snake Creek fold nappe, but had not yet equilibrated into a horizontal plane (Fig. 3.10) at the onset of Dc2. It is speculated here that isotherms, whose outcrop pattern reflect a Dc1 structure, but whose minerals are grown synchronous with (early-)Dc2, indicate a very brief time-interval between Dc1 and Dc2. Indeed, the semi-analogous saw-tooth thermal perturbations of thrust nappes with a periodicity of Lt (thickness of the thrust) decay with a time constant of $L_t^2/\pi^2\kappa$ ($\kappa$ is thermal diffusivity), so for e.g. $L_t=5$km and $\kappa=10^{-6}$m²/s, the time constant is only $\approx 80000$ years. The time-gap between Dc1 and Dc2, then, must have been very narrow, and in fact, for all reasonable strain rates and strains, Dc1 and Dc2 would have been indistinguishable, i.e. would have been one progressive deformation event instead of two discrete pulses of deformation. This contradicts the
66±25 Ma, which Page & Bell (1986) geochronologically deduced for the time-gap between D_{c1} and D_{c2} in the Sybella Granite area to the west. Alternatively, the sillimanite isograd in the Snake Creek Anticline area might reflect an as yet unrecognized, large D_{c2} structure. More integrated metamorphic/structural maps are needed to test either hypothesis.

Fig. 3.10 Three stages in the re-equilibration of an isotherm, to illustrate the offset of the sillimanite domain in respect to the axis of the Snake Creek Anticline. As a simplification the influence of pressure on the growth of sillimanite and the effect of erosion of the fold nappe are neglected. A: rapid development of a fold nappe, which folds the isotherms; B: peak metamorphism and blastesis; C: equilibrated isotherms.
The central SCG belt provides us with yet another field area in the Early to Middle Proterozoic mobile belts of northern Australian, where crustal thickening is associated with prograde low-P facies metamorphism. In order to improve our understanding of the history of the inlier, a more detailed, and integrated, structural, sedimentological, petrological and geochronological database must be compiled. The inverse approach, numerical modelling of tectono-thermal events, can put constraints on various suggested tectonic histories of the inlier. Results of a modelling study in which various tectono-thermal events were simulated (all leading to higher than normal T/P ratios), using a one-dimensional, finite difference numerical code, will be discussed in Chapter 4.
CHAPTER 4

THE SYNCHRONISM OF CRUSTAL THICKENING AND LOW-P FACIES METAMORPHISM IN THE MOUNT ISA INLIER: FAST CONVECTIVE THINNING OF MANTLE LITHOSPHERE DURING CRUSTAL THICKENING

LOOSVELD

Accepted by "Tectonophysics"; here modified for thesis
Hercyno-type orogens can be Phanerozoic, Proterozoic and Archaean!
THE SYNCHRONISM OF CRUSTAL THICKENING AND 
LOW-P FACIES METAMORPHISM IN THE MOUNT ISA 
INLIER

fast convective thinning of mantle lithosphere during crustal 
thickening

4.1 INTRODUCTION

In the previous chapter, the simultaneous development of low-P facies 
metamorphism and a phase of crustal thickening in the central Soldiers Cap Group belt 
was demonstrated. A direct effect of crustal thickening, however, must be the general 
increase of P/T ratios (resulting, in many orogenic situations, in high-P facies 
metamorphism). If moderately high metamorphic gradients could only be deduced for 
certain structural positions in the Mount Isa Inlier, like those high in a nappe sheet, the 
anomalously high temperatures could be explained by fast erosion (decompression) of the 
thickened crust and/or by the heating induced by the overlying nappe sheets (the "cover 
effect" of Davy & Gillet, 1986). Moderately high metamorphic gradients, however, are 
not restricted to these structural positions in the Mount Isa Inlier (and the other Early to 
Middle Proterozoic mobile belts of northern Australia, Fig. 1.1). Apart from prevailing 
throughout the SCG, other examples in the Mount Isa Inlier are given by Wilson (1973), 
Hill et al. (1975), Derrick et al. (1977c), Derrick (1980), Hamilton (1985), Oliver & Wall 
(1987), Reinhardt & Hamilton (in press) and Oliver et al. (in prep.). A remaining 
problem then is to find a heat source which could explain the regional steep metamorphic 
gradients during thickening of the crust.

A wide variety of fold belts have undergone regional low-P facies metamorphism, 
similar to that of the Mount Isa Inlier. Examples of this type of metamorphism can be 
found in the "Hercynotype" orogenic belts of Zwart (1967), but more commonly in 
Archean (granite-greenstone) terranes (Grambling, 1981; Green, 1981) and Proterozoic 
fold belts (Grambling, 1981; Lambert, 1983). This secular trend seems to reflect the 
overall cooling of the Earth through geological time. England & Richardson (1977) and 
England & Bickle (1984), however, pointed out that peak metamorphic conditions reflect 
points on transient and polychronous metamorphic gradients rather than on steady-state 
conductive geothermal gradients. Thus, the apparent time-dependence of peak 
metamorphic conditions should be attributed to erosion-rate dependent thermal relaxation 
of a tectonically thickened crust. Radioactive selfheating in a slowly eroding thickened 
crust may lead to overprinting of early high-pressure facies assemblages by moderate- to 
low-P facies assemblages (e.g. the overprinting of the Eo-Alpine high-P facies 
metamorphic event by the Lépontine moderate-P, moderate-T event in the European 
Alps), and, depending on reaction kinetics, even to a (near) total disappearance of the
early high-P facies assemblages. The deeper in the crust, the later peak metamorphic conditions are attained, and the more effective is the thermal relaxation (heating). England & Bickle (1984) concluded that Archaean continental thermal regimes were similar to those at present.

This hypothesis, however, does not comprehensively address the problem of the lack of relict high-P facies assemblages in the Precambrian as distinct from their Phanerozoic counterparts. In the Western Alps e.g., eclogites, glaucophane ± jadeite schists, and kyanite-bearing rocks, although overprinted by the Lépontine moderate-T, moderate-P metamorphism, are in places still recognizable (Frey et al., 1974). P-T-t paths are clockwise in the P-T field (Fig. 4.1). If tectonic and thermal regimes of the Precambrian fold belts were similar to those of the Phanerozoic, or more specifically, to those of the European Alps, the Precambrian P-T-t trajectories should resemble their Alpine counterparts and also be clockwise. Such clockwise P-T-t paths are generated in virtually all numerical models, which simulate crustal or lithospheric thickening (a.o. England & Thompson, 1984; Davy & Gillet, 1986). With certain favourable parameterisations in these models, geotherms will transect the low-P facies (andalusite-sillimanite) field for some period of time. They can only do so, however, during post-thickening decompression (−ΔP, ±ΔT), unless a combination of extreme parameters is assumed, e.g. an initial geotherm, which already transects the andalusite-sillimanite field, very low strain rates, low conductivities, and a high radioactive heat generation.

In the northern Australian Early to Middle Proterozoic inliers (Fig. 1.1), however, the regional, low-P facies metamorphism is prograde and takes place during compression (+ΔP, +ΔT). In these fold belts the early, low-P facies, prograde metamorphism was succeeded by a phase of cooling. Cooling was isobaric to slightly decompressive, as in the Mary Kathleen Fold Belt in the Mount Isa Inlier (Reinhardt & Hamilton, in press), the Arunta Inlier (Warren, 1983), and the Broken Hill Block (Phillips & Wall, 1981; Hobbs et al., 1984), or compressive at first, as in the Olary Block/Willyama Province (Clarke et al., 1987). The resulting P-T-t paths are anti-clockwise (Fig. 4.1). Such P-T-t paths have been less frequently constructed for other terranes, but Hensen (1987) using data from Arima & Barnett (1984) also deduced an anti-clockwise P-T-t path for the late Archaean Pikwitonei granulite terrane, Manitoba. Similar anti-clockwise P-T-t paths have been suggested by Bohlen (1987) for Precambrian granulite terranes as e.g. those in the Adirondacks (N.Y.), Bamble (Norway), Namaqualand (South Africa), Southern India and West Uusimaa (Finland). (Note that as yet no P-T-t paths have been reconstructed for the SCG; the low-P facies metamorphism has been described in detail, but not the isobaric cooling. In fact, kyanite has not been reported from the SCG).
Fig. 4.1 P-T diagram showing three steady-state, conductive geotherms, transecting the kyanite field. All three have a radioactive heat production (A) distribution defined by $A = A_0 e^{-D}$ (D is the length scale), a conductivity of 3 W/Km, and a lithospheric length of 100km; $D_{\text{curve } 1} = 10$ km, $D_{\text{curve } 2} = 15$ km, $D_{\text{curve } 3} = 15$ km; $A_{0,\text{curve } 1} = 2.5$ $\mu$W/m$^3$, $A_{0,\text{curve } 2} = 2$ $\mu$W/m$^3$, $A_{0,\text{curve } 3} = 2$ $\mu$W/m$^3$. Curve 3 additionally reflects a 4 km thick granite between 11 and 15 km depth with a radioactive heat production of 4.1 $\mu$W/m$^3$.

The Al$_2$SiO$_5$ triplepoint and respective stability fields are after Holdaway (1971). P-T-t trajectory "A", characterized by prograde high-P facies metamorphism and retrograde mod.-Tlmod.-P metamorphism, represents a clockwise, "Alpino-type" metamorphic history. Trajectories "B", "C", and "D" are (parts of) anti-clockwise P-T-t paths, which characterize the metamorphic history of the northern Australian Early to Middle Proterozoic fold belts (respectively after Reinhardt & Hamilton, in press; Clarke et al., 1987; and Warren, 1983): the reaction series andalusite $\rightarrow$ sillimanite (prograde) $\rightarrow$ kyanite (retrograde) is common also elsewhere (e.g. Gemuts, 1971, for the Halls Creek Mobile Belt). The cooling segment of these anti-clockwise P-T-t paths can be slightly decompressive, isobaric or slightly compressive.

Note: Abbreviations are defined in Appendix 2.
In this chapter, I will first, in a qualitative way, explore the main tectono-thermal scenarios to which regional low-P facies metamorphism has been attributed (§4.2). Then (§4.3), in order to obtain physically realistic constraints, the same scenarios will be modelled numerically, using a one-dimensional, finite-difference code (technical details of which are discussed in appendices 2, 3, 4, 5 & 6). The thermal effects of lithospheric thinning, magmatism, crust-mantle detachment and crustal thickening, as well as some combinations thereof, will be simulated. In general, there are two possible models for regional low-P facies metamorphism during crustal thickening: (i) extreme elevation of isotherms immediately before the thickening event, due to e.g. lithospheric extension, magmatic activity, or crust-mantle delamination, and (ii) crustal thickening accompanied by convective thinning of the mantle lithosphere. Only the latter possibility results in prograde low-P facies metamorphism during crustal thickening. In §4.4, the two models are compared with the existing geological database of the Mount Isa Inlier (and more briefly, the Willyama Province, Fig. 1.1). Assuming the geological database for the inlier is reliable, the former possibility can be eliminated, and crustal thickening accompanied by convective thinning of the mantle lithosphere must be the preferred solution. The results may be applicable to all Early to Mid-Proterozoic fold belts of northern Australia (Fig. 1.1), and to prograde low-P metamorphic terranes in general.

4.2 TECTONIC MECHANISMS THAT LEAD TO LOW-P FACIES METAMORPHISM

In order to explain the anomalously low P/T ratios and the subsequent isobaric cooling, one first needs to investigate three obviously possible mechanisms (Fig. 4.2B, C, D): (1) syn- to slightly pre-metamorphic lithospheric extension (McKenzie, 1978; Wickham & Oxburgh, 1985, 1987), (2) various magmatic events (Wells, 1980; Bohlen, 1987) and (3) various modes of crust-mantle detachment coupled with the upwelling of hot asthenospheric material to the base of the crust (Bird, 1978, 1979; Bird & Baumgardner, 1981; Houseman et al., 1981; Lister et al., in press). As the prograde metamorphism is contemporaneous with a phase of crustal thickening, I will, as a fourth possibility, discuss the thermal effects of this thickening, and in particular crustal thickening accompanied by an anomalously high basal heat flow, caused by convective thinning of the mantle lithosphere (Fig. 4.2E, F), as it was argued for by Houseman et al. (1981).

A mechanism not mentioned above, and not modelled here, is the late stage extension of a (thermally partially equilibrated) compressional event. Such late extension can be explained in terms of a) the gravitational instability of a topographically elevated rockmass (Frank, 1972; Artyushkov, 1973; England, 1982; Houseman & England,
1986; Platt, 1986; Royden & Burchfiel, 1987), b) the replacement of relatively strong mantle material by weaker continental crust (Vink et al., 1984; Lynch & Morgan, 1987), and c) possible heating and softening of the thickened crust by radioactive selfheating (Glazner & Bartley, 1985). Late extension is well documented in most major compressional terranes (e.g. Alps: Selverstone, 1985, 1988; Platt, 1986; Mancktelow, 1985; Norwegian Caledonides: Norton, 1986; Carpathian Mountains: Burchfiel & Royden, 1982; Andes: Suárez et al., 1983; Basin and Range, U.S.A.: Eaton, 1982; Coney & Harms, 1984; Appalachians: Snoke et al., 1980; Tibet: Molnar & Tapponnier, 1978; England, 1982; Lachlan Fold Belt, Australia: Fergusson et al., 1979). It might result in abnormally high T/P ratios, and, thus, subsequently to isobaric cooling (figure 7c of Thompson & England, 1984). This mechanism, however, is not pursued here, because in the northern Australian Early to Middle Proterozoic fold belts no evidence for such late stage extension has been reported, and P-T-t paths here are anti-clockwise in contrast to the clockwise P-T-t paths resulting from tectonic denudation, which follows crustal thickening. Indeed, the low-P facies metamorphism is clearly synchronous with crustal shortening (Hobbs et al., 1984; Etheridge et al., 1987b; Loosveld, Chapter 3).

It must be emphasized that much geological groundwork, in particular sedimentological, petrological and structural, remains to be done in the Australian Proterozoic fold belts. Most conclusions in this chapter are based and depend heavily on the as yet limited geological database, and are correspondingly speculative.

Fig. 4.2 The various tectonic models tested in this chapter (B, C, D, E & F), which all add heat to the upper lithosphere. Models B, C, D, E and F correspond respectively to numerical experiment series la, lb, lc, 2a and 2b (see also Table 4.1). The situations B, C and D are pre-metamorphic (t1), E and F are syn-metamorphic (t2). M: Moho; L: lithosphere; A: asthenosphere.

A: The steady-state conductive geotherm within an undeformed lithosphere at t=0.
B: The geotherm within an undisturbed crust and abruptly thinned mantle lithosphere.
C: The geotherm within an undisturbed crust, which is underlain by asthenospheric material. This situation is approximately similar to that of an extremely thinned mantle lithosphere.
D: The geotherm within a crust, which is intruded by a "recumbent" granite, and within a thinned mantle lithosphere.
E: The geotherm within an abruptly thickened crust, with either thickening by thrusting in the upper crust and homogeneous below that, or entirely homogeneous. The complete mantle lithosphere has been replaced by hot asthenospheric material (the Houseman et al., 1981 model).
F: As "E", but with a mid-crustal granite.
4. Numerical modelling

4.2.1 Metamorphism due to lithospheric extension

Abnormally steep metamorphic gradients, affecting large areas, are often explained by crustal/lithospheric thinning and/or addition of (extension-related pressure relief) mantle melts to the lower crust (McKenzie, 1978; Le Pichon & Sibuet, 1981; Le Pichon et al., 1982; Wickham & Oxburgh, 1985, 1987; Sandiford & Powell, 1986). McKenzie (1978) quantified the relaxation of temperatures after uniform stretching of the lithosphere (pure-shear extension). Extension can lead to asthenospheric diapirism (or vice versa), and hence to partial melting of the lower crust, the emplacement of granodiorites in the middle crust and to the generation of a condensed series of isograds. Heat flow studies of modern rift zones (Bridwell & Potzick, 1981; Lachenbruch, 1979; Lachenbruch & Sass, 1977; Mohr, 1982; Morgan, 1982) agree well with this model.

In contrast to the classical model of symmetrical and uniform stretching of the lithosphere (the "pure-shear" model), the recognition of major asymmetrical detachment fault/shear zones in the Basin and Range province (Davis et al., 1980; Wernicke, 1981, 1985; Wernicke & Burchfiel, 1982; Lister et al., 1986a; Lister & Davis, in press) led to the proposal of various asymmetric detachment models for continental extension, the "simple shear" models (Wernicke, 1985; Bosworth, 1987; Lister et al., 1986a, in press). These explain asymmetric topographic expressions, and, more importantly here, the geographic offset between areas of active (supra-)crustal extension and those affected by the highest geothermal gradients, and can therefore provide an explanation for the local absence of extensional structures in rocks characterized by low-P facies assemblages.

Both models of extension, however, have their limitations. The thermal relaxation of an instantaneously thinned lithosphere is an exponential decay process with a time constant of approximately 60 Ma. (McKenzie, 1978; Jarvis & McKenzie, 1980; Voorhoeve & Houseman, 1986), if heat is transferred by conduction only. Faster relaxation is possible if the advection of fluids plays a role. England & Thompson (1984), therefore, concluded that the thermal profiles of passive continental margins, compressed 60 Ma. after their formation, would not differ noticeably from those of resulting from compression of steady-state geotherms. Additionally, Pitman & Andrews (1985) even argued that, in small basins, thermal relaxation by lateral heat flow may largely coincide with the rifting period. For extension by means of a simple detachment zone, the lithospheric heat input is by approximately a factor two smaller than for the pure-shear McKenzie model (Fig. 4.3, and Voorhoeve & Houseman, 1988); thus, the thermal perturbation is also smaller. In the Mount Isa Inlier, this poses a formidable problem, because, although the geochronological dates on the three extensional events and on the prograde metamorphic event are widely bracketed, no extensional structures
L: Lithosphere  
A: Asthenosphere  
crust-mantle boundary at 35 km  
stippled area represents amount of heat added to lithosphere and topmost asthenosphere

Fig. 4.3 Various modes of instantaneous extension. The crust-mantle and lithosphere-asthenosphere boundaries are shown for all modes. The crosshatched area between the steady-state geotherm and the geotherm immediately after extension (bold line) represents the amount of heat added to the lithosphere and topmost asthenosphere. Radioactive heat production within the lithosphere is left out here, so the steady-state conductive geotherm is linear. "L": lithosphere; "A": asthenosphere. The results of the thermal modelling of modes A to F are given in respectively Fig. 4.5A to F.

A: 100% homogeneous and instantaneous pure shear thinning of the complete lithosphere (McKenzie-model, 1978).

B: 100% homogeneous and instantaneous pure shear thinning of the mantle lithosphere (Lister et al.-model, in press).

C: 300% homogeneous and instantaneous pure shear thinning of the mantle lithosphere (Lister et al.-model, in press).

D: instantaneous simple shear thinning of the mantle lithosphere by means of a planar, dipping, throughgoing detachment horizon, with the detachment here at the base of the crust (Voorhoeve & Houseman-model, 1988).

E: instantaneous simple shear thinning of the mantle lithosphere by means of a planar, dipping, throughgoing detachment horizon, with the detachment here in the mid-crust (Voorhoeve & Houseman-model, 1988).

F: complete excision of the mantle lithosphere, i.e. crust-mantle delamination and replacement of the mantle lithosphere by hot asthenospheric material (Lister et al.-model, in press).
have been recognized for the >100 Ma. time interval prior to prograde metamorphism (De$_3$ at 1670$^{+21}_{-17}$ Ma.; De$_2$ and low-P facies metamorphism at 1544±12 Ma.; Chapter 2).

A direct link between the extensional events and the prograde low-P facies metamorphism is therefore doubtful, and one may be tempted not to pursue this line of research here. In general, however, if the time-gap between an extensional event and metamorphism is less than 60 Ma., as it e.g. is in the 1860 Ma. old Barramundi Orogeny (Etheridge et al., 1987b), crustal extension would not only explain the low-P facies character of that metamorphism (although not necessarily andalusite-sillimanite facies), but also the subsequent isobaric cooling, which is petrologically deduced in many of the northern Australian Early to Middle Proterozoic fold belts.

Another possibility is that, as yet unrecognized, late (i.e. immediately pre-metamorphic) asymmetric rifts are locally developed. The thermal evolution of the tectonically undisturbed crustal areas adjacent to such a late supra-crustal rift can be affected by (i) lateral heat flow, aided by the generation of small-scale convection cells in the asthenosphere under a rift margin due to the steep horizontal temperature gradients (Buck, 1986), and (ii) the thinning of the mantle lithosphere under the undisturbed crust, either by pure shear as in the asymmetric extensional model of Lister et al. (1986a, in press), or by simple shear as in the asymmetric extensional models of Wernicke (1985) and Voorhoeve & Houseman (1988). All asymmetric extension models predict a horizontal offset between that part of the lithosphere, which is most attenuated, and that part of the upper crust, in which rift structures are developed: anomalously steep geothermal gradients, due to the thinning of the mantle lithosphere (± lower crust) and the possible detachment of the upper mantle, will affect areas of relatively undisturbed upper crust (the hanging wall, i.e. the "upper plate" of Lister et al., 1986a, in press). This upper plate situation will be simulated here, omitting the lateral heat flow complication.

4.2.2 Heating by intrusion of magmas

Crustal thickening by magmatic accretion, i.e. both under- and over-accretion, will result in very high relative temperatures in all levels of the crust (Wells, 1980), and temporarily very steep geothermal gradients above the intrusion (100-500°C/kbar, Bohlen, 1987). Rocks above the intruded material will heat isobarically, and subsequently, during the thermal relaxation, cool isobarically. Highly differentiated granites can (even after emplacement and cooling) continue to play an important role in the thermal balance of an area, because they are commonly enriched in heat producing elements (HPE). In this study, we will calculate the anomalously high heat production in the pre-metamorphic granites of the Mount Isa Inlier.
Wyborn et al. (in press) argue that the granites of the Mount Isa Inlier are the result of a two-step differentiation. As magmatic events represent a transient stage in the re-equilibration of a highly perturbed thermal (or pressure or chemical) gradient, these granites must be the result of two successive thermal perturbations. Discussing the first thermal perturbation, Wyborn et al. (in press) mention the possibility of both extensional and compressional tectonic events, which can lead to mantle melting, and, mainly because primary magmas from the mantle at crustal depths are denser than crustal material (Herzberg et al., 1983), to underplating at the base of the crust. These underplates could then, during a second perturbation, be the source of the major I-type batholiths of the Mount Isa Inlier.

As intrusive activity does not reflect steady-state tectono-thermal conditions, but instead results from a highly perturbed geotherm (and is thus a second-order process!), it is here simulated with either extension or contraction.

4.2.3 Crust-mantle detachment

In most cases of underplating, exchange of cold, dense (3300kg/m$^3$; Molnar & Gray, 1979) lithosphere by hot, less dense (3280kg/m$^3$ according to Molnar & Gray, 1979; 3220kg/m$^3$ according to Hargraves, 1981) asthenosphere under the base of the crust, plays a role (Wyborn et al., in press). Such substitution can be explained in four ways: (1) by conductive thinning of the mantle lithosphere, due to a heat flow perturbation (thermal "plume"; Crough & Thompson, 1976); (2) by "delamination", i.e. the vertical separation of crust and upper mantle by the bending of a coherent slab of dense mantle lithosphere, whose tip sinks into the less dense and hotter asthenosphere (Bird, 1978, 1979; Bird & Baumgardner, 1981); (3) by "convective thinning" of the mantle lithosphere during crustal thickening (Withjack, 1979; Houseman et al., 1981); (4) by juxtaposition of hot asthenospheric material in the footwall of a major throughgoing (extensional) detachment zone, and cooler lithospheric or crustal material in its hanging wall.

Crust-mantle delamination (2) and convective thinning (3) are based on the gravitationally unstable layering of the continental lithosphere: relatively dense mantle lithosphere overlies less dense asthenosphere. Thus, regardless of the style of crustal deformation, a deflection from the horizontal on the crust-mantle boundary will tend to grow until the lowest potential energy level is reached, i.e. the density contrast is reversed. Independent of the mode of substitution of the mantle lithosphere by asthenospheric material, it invariably leads to anomalously high temperatures in the crust. Etheridge et al. (1987b) have proposed a working hypothesis for the 1860 Ma. old
Barramundi Orogeny in the northern Australian Early to Middle Proterozoic fold belts, which incorporates extension-triggered crust-mantle delamination. The 1550 Ma. old metamorphism in the Mount Isa Inlier, however, differs from the widely recognized 1860 Ma. old event in that the basin-forming tectonism is at least 100 Ma. older than the low-P facies metamorphism. Here, mantle lithosphere substitution will be incorporated in some of the extensional and compressional models.

4.2.4 Crustal thickening

It is argued in this chapter that, although extension and magmatism, when immediately predating crustal thickening, can result in low-P facies metamorphism and subsequent isobaric cooling, these models do not otherwise fit the geological history of the Mount Isa Inlier. Since the prograde regional low-P facies metamorphism in the inlier is clearly coeval with a pervasive phase of crustal thickening (+ΔP, +ΔT), one is forced to re-examine the effects of thickening on the evolving temperature-depth profile. Brady (1982) modelled one-dimensional heat transfer in an orogenic setting, and summarized the factors possibly leading to andalusite stability in such an environment:

1. syn-metamorphic regional plutonism;
2. an unusually insulating sediment pile;
3. transport of hot rocks to shallow levels by tectonic movements or by rapid erosion;
4. a mantle heat flux triple the normal value;
5. a very thick radioactive crust;
6. a volatile flux equivalent to one rock volume of fluid per 5000 yrs.

Additionally, Reitan (1968a, 1968b, 1969), Graham & England (1976), Molnar et al. (1983) and Werner (1985) considered frictional heating in tectonically active areas, such as subduction and overthrust terranes. This, however, could only condense isotherms locally (and even lead to locally inverted metamorphic geotherms), but cannot lead to a widespread, significant temperature increase (England & Thompson, 1984). Additionally, there are no indications of frictional heating playing a more prominent role in the Australian Proterozoic terranes than in any other orogen characterized by a clockwise P-T-t path.

Of the six factors given by Brady (1982), there is no direct evidence for factors 1, 2, 3 and 4 in the Mount Isa Inlier. As pointed out in §4.1, crustal thickening as such will generally not lead to coeval low-P facies metamorphism. Crustal thickening will lower a rock in the pile, adiabatically, or with slight heating (ΔP positive, ΔT=0). The subsequent heating-up phase will be accompanied by isostatic uplift and consequent erosion. Possible melting in the lower crust by this thermal relaxation is therefore

With respect to factor 2 (the unusually insulating sediment pile), Jaupart & Provost (1985) considered conduction of heat across the Main Central Thrust of the Himalayas. They concluded that the location of young leucogranites, at the top of the basement sheet, can be explained by the heterogeneous vertical thermal conductivity distribution, i.e. a low conductivity in the upper sediments versus a high conductivity in the lower crystalline basement. There is no evidence, however, that a similarly insulating sequence has overlain the northern Australian Proterozoic inliers during the metamorphism.

Factor 3 (rapid upward transport) only plays a role if the upward transport is faster than thermal relaxation. As mentioned before, folds in the Mount Isa Inlier are generally tight to isoclinal and upright and the lower P/T ratios might be expected in the major antiforms. Even though andalusite-sillimanite blastesis is synchronous with this folding event, such a structurally controlled P/T ratio distribution has not been documented and thermal relaxation must be faster than deformation. Indeed, for reasonable strains and strain rates, this must be generally the case with upright folding: thermal relaxation of a lateral perturbation of periodicity \( L_f \), with \( L_f \) the half-wavelength of a fold, has a time constant of \( L_f^2/\kappa^2 \) (with \( \kappa \) the thermal diffusivity), so that e.g. for \( L_f = 10\, \text{km} \) and \( \kappa = 10^{-6}\, \text{m}^2/\text{s} \), the thermal time constant is only \( \approx 0.3\, \text{Ma} \). Widespread uplift by erosion and/or tectonic denudation do not have to be considered either as the low-P facies metamorphism is synchronous to crustal thickening. Additionally, the post-compressional segment of the P-T-t path is essentially isobaric (Reinhardt & Hamilton, in press).

As far as the mantle heat flux (factor 4) is concerned, it can be calculated that the rate of mantle heat production 1.6 Ga. ago was 20 to 40% higher than it is now (assuming \( C_0^\text{M}/C_0^\text{O} = 10^4 \); McKenzie & Weiss, 1975; Davies, 1980; Turcotte & Schubert, 1982). If none of the more conventional explanations are acceptable, channelized mantle heat flow can be an explanation for local prograde low-P facies metamorphism. An elevated, but constant basal heat flow, however, will not affect the sense of rotation of the P-T-t path (clockwise versus anti-clockwise).

Disregarding factors 1, 2, 3 and 4 leaves only radioactive selfheating and heat transfer by fluid advection processes as the most important factors. England & Thompson's (1984) and Davy & Gillet's (1986) studies on the thermal balance of orogenic zones are in this regard the most detailed. The England & Thompson model (1984) quantitatively incorporates radioactive selfheating after instantaneous
crustal/lithospheric thickening, and qualitatively discusses fluid advection. The Davy & Gillet (1986) model also omits fluid flow from the numerical code, but introduces multiple time-dependent, finite thrusting events. Both models explain low-P facies retrograde metamorphism (or rather conditions: -ΔP, ±ΔT) by the combination of slow uplift and an increased radiogenic heat supply in the thickened crust. Instantaneous thickening of a crust with a normal geothermal gradient, as modelled in the England & Thompson study, is unlikely to result in a purely prograde traverse of the andalusite-sillimanite field. This holds both for models with initial eclogite-"sinkers" (Richardson & England, 1979; England & Thompson, 1984; Thompson & Ridley, 1987) and without them. P-T-t paths from these experiments are invariably clockwise. In the Mount Isa Inlier, nevertheless, andalusite and sillimanite blastesis is prograde and compressive (Derrick et al., 1977c; Reinhardt & Hamilton, in press; Loosveld, Chapter 3), and is followed by essentially isobaric cooling (Reinhardt & Hamilton, in press): the P-T-t paths are anti-clockwise.

Deformation, however, can increase permeabilities and fluid flow can become an additional mechanism of heat transfer (Fyfe et al., 1978; Etheridge et al., 1983, 1984a; Ferry, 1984; Bickle & McKenzie, 1987). In the Mary Kathleen Fold Belt in the central Mount Isa Inlier, large-scale fluid flow has been proposed by Oliver et al. (1987) and Oliver & Wall (1987). In this study, numerical experiments simulating crustal thickening will, in simplified form, include single-pass fluid flow and its associated advection of heat (the role of convection, i.e. multi-pass fluid flow, in rocks below 3-6 km is still contentious; e.g. Etheridge et al., 1984a, versus Wood & Walther, 1986).

I will also evaluate a special case of continental convergence or thickening, which combines crustal thickening with convective thinning of the subcrustal lithosphere. This case was argued for by Houseman et al. (1981) for a lower lithosphere and asthenosphere with the same constant viscosity, and by Fleitout & Froidevaux (1982) for a stratified, viscous Newtonian lithosphere. It differs markedly from the above-mentioned conventional lithospheric thickening models, as it would add an enormous amount of heat to the base of the (thickening) crust. As the relatively cool and dense lithospheric mantle thickens, it is submerged into the underlying, hot and less dense asthenosphere. Thus, its gravitational instability is increased, and it may (partially) detach from the overlying crust (and maybe uppermost mantle lithosphere) to sink into the asthenosphere (Fig. 4.4). Houseman et al. (1981) argued in favour of the detachment of the entire mantle lithosphere from the crust, exposing the lower crust to asthenospheric temperatures (1250-1350°C).
This crustal thickening with convective thinning of the subcrustal lithosphere may be in good agreement with the various observed metamorphic phenomena of the northern Australian Early to Middle Proterozoic mobile belts like the anti-clockwise P-T-t trajectories, the simultaneity of crustal thickening and low-P facies metamorphism, the concave curvature of the metamorphic geotherm towards the temperature-axis (Ryburn et al., in prep.) and the, in places, enormous amounts of post-tectonic granites (e.g. the Williams and Naraku Batholiths). Quantitatively, the following questions have to be resolved: first, under what conditions does a combination of radiogenic selfheating in a thickening crust, advection of heat by fluid flow, and convective thinning of the mantle lithosphere lead to prograde andalusite-sillimanite blastesis? And second, under what conditions is the prograde andalusite-sillimanite stage succeeded by essentially isobaric cooling, thus giving rise to an overall anti-clockwise P-T-t path?

Fig. 4.4 Five schematic stages of progressive crustal thickening accompanied by progressive convective thinning of the mantle lithosphere. Stage 1 represents a metastable conductive lithosphere overlying convecting asthenosphere. At the onset of crustal thickening (stage 2), convection at the base of the lithosphere is enforced as the relatively cold and dense lithospheric root of the incipient mountain belt is submerged into relatively hot and less dense asthenosphere. As thickening continues, more lithospheric material is swept sideways into the downgoing plume (stage 3), until the entire (?) lithosphere is detached from the crust (stage 4). If convection of asthenospheric material continues under the base of the crust during the remaining stages of crustal thickening (stage 5), then a continuous and extremely high basal heat flow will accompany these later stages of crustal thickening. After a model by Houseman et al. (1981). Vertical scale greatly exaggerated.
4.3 RESULTS OF THE NUMERICAL EXPERIMENTS

In all reported cases of low-P facies metamorphism in the northern Australian Proterozoic fold belts, the prograde metamorphism is coeval with a period of crustal thickening. Hence, the thermal effects of any mechanism that has led to the anomalously high temperatures must have been interacting with the (initial) perturbation of the geotherm by crustal thickening. Local thickening of the entire lithosphere cannot by itself have led to the prograde low-P facies metamorphism and anti-clockwise P-T-t paths (e.g. England & Thompson, 1984). Combining the various pre- to syn-metamorphic modes of crustal heating with syn-metamorphic crustal thickening will therefore be essential. Table 4.1 lists the possible combinations.

Table 4.1 Possible combinations of events, which lead to low-P facies metamorphism. Combinations marked by an asterisk are considered possible for the Australian Proterozoic fold belts and are modelled here (explanation in text).

<table>
<thead>
<tr>
<th>Exp. series</th>
<th>pre-metamorphic</th>
<th>syn-metamorphic</th>
<th>Figure</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a*</td>
<td>extension (pure shear)</td>
<td>(compression)</td>
<td>4.2B</td>
</tr>
<tr>
<td>1a2*</td>
<td>extension (simple shear)</td>
<td>(compression)</td>
<td></td>
</tr>
<tr>
<td>1b*</td>
<td>ext. + asthenospheric upwelling</td>
<td>(compression)</td>
<td>4.2C</td>
</tr>
<tr>
<td>1c*</td>
<td>ext. + magmatism</td>
<td>(compression)</td>
<td>4.2D</td>
</tr>
<tr>
<td>1d</td>
<td>magmatism</td>
<td>(compression)</td>
<td></td>
</tr>
<tr>
<td>1e</td>
<td>asthenospheric upwelling</td>
<td>(compression)</td>
<td></td>
</tr>
<tr>
<td>1f</td>
<td>magm. + asthenospheric upwelling</td>
<td>(compression)</td>
<td></td>
</tr>
<tr>
<td>2a*</td>
<td>-</td>
<td>asthenospheric upwelling + compression</td>
<td>4.2E</td>
</tr>
<tr>
<td>2b*</td>
<td>-</td>
<td>asthenospheric upw. + magm. + compr.</td>
<td>4.2F</td>
</tr>
<tr>
<td>2c</td>
<td>-</td>
<td>magmatism + compression</td>
<td></td>
</tr>
<tr>
<td>3-20</td>
<td>variations of 1 and 2 series</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Due to geological constraints, only a few combinations need to be investigated. As thermal relaxation after thickening of the entire lithosphere generally leads to late- to post-tectonic melting in the lower crust, such magmatism will have no effect on the early- to syn-compressional geotherm, and the possibility of this combination (No. 2c) is discarded here. Also discarded are those combinations with isolated magmatism and/or asthenospheric upwelling (no. 1d, 1e & 1f), since magmatism and asthenospheric upwelling are envisaged to take place in a tectonically active setting. Here, only the thermal effect of asthenospheric upwelling is considered, regardless of the mechanics of replacement of cold lithosphere by hot asthenosphere. Possibilities 3-20 (Table 4.1) combine the components of experimental series 1 and 2, and are thus more complex.

Note: Numerical modelling was not just to explain metamorphism in the NB, Isua Index.
involving more assumptions and free parameters. Because of the complexity, and in order to evaluate isolated rather than randomly combined processes, possibilities 3-20 are also left out. This leaves an early extension with or without asthenospheric upwelling and/or magmatism, followed by compression (Nos. 1a, 1b & 1c), and compression combined with asthenospheric upwelling, with or without magmatism (Nos. 2a & 2b). These five possibilities are depicted in Fig. 4.2B-F.

The first group of numerical experiments deals with two events, an early crustal extension with optional asthenospheric upwelling and magmatism, followed by a phase of compression (Table 4.1: series 1). The second group of experiments is concerned with one tectonic event, i.e. Houseman et al.'s (1981) model, consisting of crustal thickening combined with coeval asthenospheric upwelling and optional magmatism (series 2).

In all numerical experiments, the one-dimensional conservation of energy equation is solved with a simple, one-dimensional, iterative, finite difference (Crank-Nicolson), BASIC code on a Macintosh™. The lithosphere is subdivided into 50 equally spaced depthsteps (technical details in appendices 2-7).

4.3.1 Experiment series 1: lithospheric thinning

If a metamorphic gradient is to transect the andalusite-sillimanite field, it has to be dramatically steeper than the steady-state geothermal gradient (Fig. 4.1). Whether andalusite/sillimanite-conditions are stable during a compressional event, which postdates an extensional event, will depend on such variables as the amount and style of extension, the location within the extensional setting, the time interval between extension and compression, and the ratio of the compressional strain rate to the thermal relaxation rate. Additional heat sources may result from crust-mantle delamination or magmatic intrusion. In some areas of the Mount Isa Inlier, peak metamorphism is coeval with the early stages of the penetrative compressional event (the Soldiers Cap Group: Loosveld, Chapter 3), and in that case the stretching of the geotherm due to crustal thickening may be neglected. Still, temperatures on all levels have to be increased by at least 100-200°C (Fig. 4.1). The abrupt change of the geotherm by the various modes of "instantaneous" extension is depicted in simplified form in Fig. 4.3A to F. Figures 4.5A to F give the thermal relaxation on ten depth levels associated with the situations in Fig. 4.3A to F respectively. The steady-state conductive geotherm in the experiments of Fig. 4.5 is relatively "hot", as it includes a cooled, 4 km thick HPE-enriched granitic slab between the 18 and 22 km levels (representing conditions of the Mount Isa Inlier). This steady state geotherm is obtained numerically.
A 100% extension of mantle lithosphere

B Homogenization of entire lithosphere

C 300% extension of mantle lithosphere

D Detachment at 40 km

E Detachment at 20 km

F Complete excision of mantle lithosphere
4.3.1.1 extension alone: experiment series 1a

Three modes of - instantaneous - lithospheric extension were simulated:
1. homogeneous "pure shear" thinning of the entire lithosphere after McKenzie (1978);
2. homogeneous "pure shear" thinning of the mantle lithosphere while not deforming the crust (the "upper plate" of Lister et al.'s (in press) models);
3. thinning of the lithosphere by a planar, dipping detachment horizon (Voorhoeve & Houseman, 1988).

An abrupt homogeneous, plane strain "pure shear" extension of the entire lithosphere by a factor of 2 (β=2), superposed on slightly steeper-than-normal steady-state conductive geothermal gradients, will generally result in sillimanite conditions (Fig. 4.5A). Thermal relaxation (cooling), however, will be fast, aided by the concomitant thinning of the HPE-enriched upper crust by a factor 2.

This thinning of the HPE-enriched upper crust is absent when simulating the upper plate situation of asymmetric extensional models, where a thinned mantle lithosphere underlies an undisturbed crust (Fig. 4.3B, C, D), or where the mantle lithosphere is completely excised (Fig. 4.3F). On the other hand, the heat input to the lithosphere (represented by the shaded area in Fig. 4.3) is, at the same amount of crustal extension, relatively smaller in the asymmetric extensional models compared to the pure shear lithospheric thinning (Voorhoeve & Houseman, 1988; Fig. 4.3). When adding these
competing effects, the lower crust still effectively buffers temperature increases. If thermal relaxation is attributed to conduction alone, even an "instantaneous", 300% pure-shear extension of the mantle lithosphere, immediately followed by conductive cooling, will generally not lead to sillimanite grades in the middle crust (Fig. 4.5C). Simple shear experiments (experiment series 1a2) yield results even further removed from sillimanite conditions (Fig. 4.5D), except for those situations in which the detachment zone happens to transect the middle crust (Fig. 4.5E). Loosveld & Etheridge (Appendix 8) argue that andalusite-sillimanite conditions will generally only result from lithospheric extension (without magmatism and/or delamination) if extension factors equal or exceed 3. Such large extension factors are typical for the outward parts of passive continental margins (Le Pichon et al., 1982), but are unlikely to be found in intra-continental rifts.

In many numerical experiments simulating various modes of extension, the temperature in the lower crust exceeded the solidus temperature for some time, e.g. after pure-shear mantle lithosphere extension of 300%, the temperature at the 30 km level is briefly raised to a maximum 660°C (Fig. 4.5C), so that melting and subsequent advective heat transport may take place. Pressure-release melting of the mafic upper mantle will add to the extent of magmatism and may trigger more lower crustal melting.

Whether or not andalusite-sillimanite conditions are reached will depend among others on the mode of "instantaneous" extension, but, independent of the mode, these conditions will not be stable for more than 10 to 15 Ma. (Fig. 4.5A, E, and F). As extension will not be instantaneous, T/P ratios will in fact be lower. In all modes of extension, thermal relaxation is fast; after 50 Ma., temperatures in the middle crust are only slightly higher than the original steady-state ones. Additional heat transfer mechanisms, other than conduction, would result in an even faster thermal relaxation. In appendix 8, Loosveld & Etheridge again emphasize the temporal and spatial limitations of the extensional model.

4.3.1.2 extension + delamination: experiment series 1b

Under an "upper-plate" passive margin, the mantle lithosphere can be attenuated or completely excised (Lister et al., 1986a, in press). In the latter case, hot asthenospheric material underplates the lower crust (Fig. 4.3F). This situation can be simulated using the same formulation as with experiment series 1a1 by making the pure shear extension factor within the mantle lithosphere infinitely large. Relaxation paths after such crust-mantle delamination with the upwelling of asthenospheric material to the base of the crust were calculated both with ongoing convection in the anomalously shallow asthenospheric material, and with immediate "freezing" of the asthenospheric material (respectively "hot
A

"cold mode" delamination

B

first 1 Ma "hot mode"

C

first 20 ma "hot mode"
mode" and "cold mode" crust-mantle delamination of Bird & Baumgardner, 1981). The "hot mode" is simulated by keeping the temperature at the base of the crust constant for a given time, whereas in the "cold mode" the lower temperature boundary is shifted back to the base of the pre-extension lithosphere (100-150 km) immediately after delamination.

Considering "cold mode" delamination and heat transfer in the crust by conduction only, the andalusite-sillimanite field may, depending on the steady-state conductive geotherm, conductivity and the thickness of the "crust" above the detachment horizon, be reached at the deeper crustal levels (compare Fig. 4.5F with 4.6A, and Bird, 1979); in experiment 135 (Fig. 4.5F), in which the mantle lithosphere (including the 36 km depth level) is detached, sillimanite conditions were reached at the 26 km level between 2 and 12 Ma. after the instantaneous delamination. Experiment 42 (Fig. 4.6A), on the other hand, in which the detachment horizon is at 37.5 km depth, and in which a slightly cooler steady-state, conductive geotherm was chosen, did not yield sillimanite conditions at similar depth levels. Peak temperatures in the upper crust are reached between 10 and 20 Ma. after delamination, but are not high enough to trigger sillimanite growth. In experiment 135, andalusite and sillimanite are only stable in the first 12 Ma.. This implies that the time-gap between the compressional event, i.e. the actual time of andalusite/sillimanite growth, and this "cold mode" delamination could not have exceeded 12 Ma.. Considering the progressive, vertical stretching of isotherms during crustal thickening, "cold mode" delamination cannot explain the synchronicity of crustal thickening and low-P facies metamorphism.

\[ T = 281.281e^{215} + 7.13x, \text{ with } x \text{ in km, for } x \leq 36 \text{ km, and } T = 1350^\circ C \text{ (in A) and } 1300^\circ C \text{ (in B and C) for } 39 \leq x < 150 \text{ km. } T_{\text{max}} \text{ (at 150 km)} = 1350^\circ C; K = 2Wm^{-1}K^{-1}; \kappa = 10^{-6}m^2s^{-1}; D = 15km, A_o = 2.5\mu Wm^{-3}. \text{ The 4 horizontal crosslines represent the temperatures at which either sillimanite or andalusite becomes stable at respectively the 9, 15, 21 and 27 km level. The dotted fields in B and C represent the time/depth domain in which andalusite or sillimanite is stable. Transfer of heat in the crust after } t=0 \text{ is by conduction only.}

A: cold mode crust-mantle delamination. No andalusite/sillimanite blastesis.

B: hot mode crust-mantle delamination for 1 Ma., i.e. the temperature at the base of the crust is kept at 1300°C for 1 Ma., thus simulating convection of heat in the newly upwelled asthenospheric material at the base of the crust.

C: hot mode crust-mantle delamination for 20 Ma., i.e. the temperature at the base of the crust is kept at 1350°C for 20 Ma.
To obtain higher crustal temperatures for longer time intervals, another heat source has to be sought, either some form of "hot mode" upwelling of the asthenosphere, or immediate differentiation of the magmatic products leading to upper crustal granitic slabs enriched in heat-producing-elements (HPE). Hot mode delamination will lead to the expansion of andalusite-sillimanite conditions to higher crustal levels. In Fig. 4.6, three experiments with respectively 0, 1 and 20 Ma. hot mode underplating are depicted. This figure shows, as expected, that hot mode delamination would be a very powerful heat source at the base of the crust.

4.3.1.3 extension + magmatic events: experiment series 1c

The notion that heat might advect to the upper crust by rising magma was briefly mentioned by Wickham & Oxburgh (1985); the thermal perturbation of a mafic magma emplaced at a depth of 15 km and with a thickness of 10 km would, however, only be short-lived. Although true for mafic magmas, such a consideration does not hold for fractionated felsic magmas, because it neglects the possible progressive enrichment of heat producing elements (HPE) with magmatic differentiation and partial melting. It is not uncommon that a mid-crustal or upper-crustal slab is replaced by an igneous slab with a radioactive heat production an order of magnitude higher. Such HPE-enriched granites are powerful and permanent heat sources. Intrusion of such granites in the middle or upper crust has two thermal effects: the first one the transient cooling of the magma and concomitant heating of the country rock (contact metamorphism), the other effect is brought about by the HPE-distribution anomaly, which will permanently cause anomalously high temperatures.

The abundance of HPE-enriched granites in the Mount Isa Inlier (and other Australian Proterozoic fold belts; Wybom et al., 1987) justified a closer examination of the thermal effects of granitic intrusions. Seismic, heat flow and gravity data have led to the concept of "recumbent" batholithic masses (Hamilton & Myers, 1967). Thickness estimates of the Sierra Nevada Batholith vary between 5 and 15 km. Here, we use 3, 4, 6, 9 and 12 km thick slabs of granite, and internal heat productions within these granites ($A_{gr}$) of 2.5, 4.1, 4.75, 5.0, and 7.5μW/m². The mineralogy, textures and setting of the Mount Isa granites indicate that they have intruded the upper crust (L.A.I. Wybom, pers. comm., 1987). Taking into account their regional metamorphic grades, they must have subsequently been submerged to mid-crustal levels, either tectonically or by sedimentary loading. I have assumed the top of the granites at 18 km. Figure 4.7 shows the thermal effect of the presence of such a differentiated granite in the rock pile. From Fig. 4.7, it can be concluded that temperatures around thick and highly HPE-enriched granitoids - if the temperature in such HPE-enriched magmas could fall under the solidus temperature in
Fig. 4.7  The long-term thermal effect of the intrusion of a HPE-enriched granite on the overlying country rocks. In A1 to A4, 
$T_{t=0} = T_{t=-1}$ is plotted versus the distance to the top of the granite for various $A_{gr}$. The 
thickness of the granitic slab is in A1: 3km; in 
A2: 6km; in A3: 9km; and in A4: 12km. In 
B1 to B4, the same family of curves is plotted, 
but differently grouped: $T_{t=0} = T_{t=-1}$ is plotted 
versus the distance to the top of the granite for 
various thicknesses of the granitic slab. The 
radiogenic heat production in the granite is: B1: 
$A_{gr}=2.5\mu W/m^3$; B2: $A_{gr}=4.75\mu W/m^3$; B3: 
$A_{gr}=5.0\mu W/m^3$; B4: $A_{gr}=7.5\mu W/m^3$. The 
top left region of the plots represents 
temperatures above the solidus, and is therefore 
unstable. Experiment parameters: $L=150km$; 
the steady-state conductive geotherm before 
intrusion is defined by $T_{max} = 1350^\circ C$, $A_0 =$ 
$2.5\mu W/m^3$, $D=15km$ and $K= 2W/Km$. The 
top of the intrusion is in each case at 18km 
depth.  

$T_t = T_{t=-1}$

+ increasing distance to top of granite
the first place - would be raised above melting point. Mixing of granitoid magma with country rock melt, then will reduce the HPE-concentration. Figure 4.1 shows the long-term change of a geotherm after emplacement of a 4 km thick granite between the 18 and 22 km levels (with $A_{gr}$=4.1$\mu$W/m$^3$).

From experiment series 1, it can be concluded that low-P facies metamorphism can only be explained by transient thermal pulses caused by extensional events, magmatism or cold mode delamination, if these events immediately predate the metamorphism. Lower and mid-crustal rocks will for a limited time remain anomalously hot: in the case of magmas in the order of a few million years, and in the case of extension and/or cold mode delamination a few tens of millions years at the most. For the low-P facies metamorphism to be prograde (+$\Delta T$), the time-gap has to be even narrower (at the most $\leq$10 Ma.; Fig. 4.5). Prograde and compressive (+$\Delta T$, +$\Delta P$) low-P facies metamorphism cannot be explained by any extensional/magmatic event. To obtain higher temperatures for longer periods, either some form of hot mode delamination, or the presence of large HPE-enriched granitoids in the crust has to be invoked.

4.3.2 Experiment series 2: crustal thickening

The first segment of the P-T-t path deduced for the Mount Isa Inlier shows compression with slightly decreasing P/T ratios (Reinhardt & Hamilton, in press; Fig. 4.1). As thermal relaxation by conduction of a thickening lithosphere is for all reasonable parameters slower than the lithospheric thickening itself, P/T ratios should, however, increase during thickening rather than decrease. One must conclude that either heat is transferred by a more effective process than conduction alone (e.g. advection of a single-pass fluid flow), or that crustal thickening in the inlier (and many others) is associated with an anomalous heat pulse.

4.3.2.1 the effect of single-pass fluid flow

The effective fluid flux, W, is normally deduced from D'Arcy's Law, or calculated from geochemically inferred fluid/rock ratios, which will give a minimum. As, at present, there are few reliable estimates on fluid/rock ratios and original rock permeabilities (in Early to Mid-Proterozoic fold belts), I have used an alternative approach. Walther & Orville (1982) calculated the amount of post-diagenetic devolitalization products during regional metamorphism of an average pelite as being 12 vol%. Assuming a thickness, h, in meters of pelitic material, and assuming a constant fluid flow rate during deformation, and disregarding the fact that rocks undergoing the transition from andalusite-bearing assemblages to sillimanite-bearing assemblages would
already have been relatively dry, "W", the effective fluid flow (in m/s) above this pelitic layer is approximated as \((0.12h)/(\text{time of deformation in seconds})\). This method is not only simple to incorporate in the numerical code, but it also ensures that no unrealistic amount of fluids is being withdrawn from the crust. The effective fluid flows obtained in this way fall within the wide range of fluid flows calculated with D'Arcy's Law (Etheridge et al., 1984a). The result of constant fluid flow rates during deformation is a constant extra heat input on any crustal level above the pelitic layer during deformation. With higher strain rates, the difference between a sudden devolitalization and constant devolitalization rates over the deformation period will be smaller.

By generously assuming "h" as 20 km, the final amount of expelled fluid has a column thickness of \(\mathbf{24}\) km. In Fig. 4.8, the effect of the steady expulsion of this

![Fig. 4.8](image)

**Fig. 4.8** The effect of upwards directed single-pass fluid flow during progressive, homogeneous crustal thickening, with a constant strain rate. From the 4 P-T-t paths shown, number 1 and 3 result from experiments incorporating a fluid advection component; number 2 and 4 result from experiments without the fluid flow, and are plotted for comparison. In all experiments \(T_{\text{max}}=1350^\circ\text{C}; K=2\text{W/Km}; \kappa=10^{-6}\text{m}^2\text{s}^{-1}; D=15\text{km}; A_0=2.5\mu\text{Wm}^{-3}; f_{\text{end}}=1.6; \) the initial lithospheric thickness is 150km; the initial crustal thickness is 35 km; \(\rho_{\text{crust}}=2700\text{kg/m}^3; \rho_{\text{subcrust}}=3300\text{kg/m}^3\). Strain rate, \(\dot{\varepsilon}\), of experiment 1 and 2 was \(10^{-14}\text{s}^{-1}\) (duration of thickening is 3 Ma.); \(\dot{\varepsilon}\) of experiment 3 and 4 was \(10^{-15}\text{s}^{-1}\) (duration of thickening is 30 Ma.). At the end of thickening, thermal relaxation was by conduction only. No decompression accompanied thermal relaxation after thickening.
amount of fluid during two experiments (curves 1 and 3), involving constant strain-rate deformation is shown and compared with two identical experiments (curves 2 & 4), without the fluid flow. One may conclude that the advection-of-heat effect for single-pass fluid flow is only minor.

4.3.2.2 crustal thickening and associated convective thinning of the mantle lithosphere

The only remaining model that can explain the synchroneity of low-P facies metamorphism and crustal thickening is Houseman et al.'s (1981) model, which combines crustal thickening with convective thinning of the mantle lithosphere (Fig. 4.4). This model can result in the prograde transection of the andalusite-sillimanite field during crustal thickening, if thickening is slow relative to fast (=early) convective thinning of the upper mantle. Then, the condensing of isotherms in the crust due to increased basal heat flow will compete with the stretching of isotherms, caused by the (progressive) crustal thickening. Hence, the temperature on a certain level may rise, and the overall geothermal gradient may steepen. Subsequently, at the end of convection in the upwelled asthenosphere (assumed to be at the end of crustal thickening), when the basal heat flow decreases again, the crust may cool isobarically or slightly decompressively. Thus, an overall anti-clockwise P-T-t path may result.

Houseman et al.'s (1981) study formulated the relation between instantaneous thickening, \( f \), and the time, \( t_o \), of the peak of kinetic energy, which coincides with the time of detachment of the thermal boundary layer from the upper rigid layer: \( \log(t_o) \) is inversely proportional to \( f \). The time of the peak of kinetic energy, \( t_o \), ranged in their experiments between 4 and 88 Ma., with the bulk of the runs between 10-30 Ma.. This probably is in the order of the time-scale of a crustal thickening event. Houseman et al. (1981) did not include experimental runs with a progressive deformation, but concluded that the base of the crust may well be subjected to asthenospheric temperatures during the (later) stages of thickening.

The viability of anti-clockwise P-T-t paths as a result of Houseman et al.'s (1981) model is tested here by a few experiments with widely varying parameters (experiment series 2). Crustal thickening was approximated as being homogeneous throughout the crust. The finite difference grid was fixed to the medium and thus deformed with thickening (the change in timestep-spacing being quadratically proportional to the change in depthstep-spacing). Strain rates were kept constant; values between 10^{-14.3} and 10^{-15.3}s^{-1} were used. The finite amount of thickening was by a factor 1.6 (total time of thickening therefore varies between 3 and 30 Ma.). Convective thinning of the mantle
Fig. 4.9  *P*-*T*-*t* paths resulting from progressive, homogeneous crustal thickening accompanied by convective thinning of the mantle lithosphere. The strain rate was constant in each given experiment. The progressive substitution of the mantle lithosphere by asthenospheric material was approximated by an abrupt shift of the lower boundary condition (*T*=1300°C) to the base of the crust at *t*<sub>det</sub>. The temperature (1300°C) at the base of the crust was kept constant during crustal thickening. At the end of thickening, the lower boundary condition was abruptly shifted back to its original position (the base of the pre-thickening lithosphere). Heat transfer in the crust was by conduction and by an upwards directed single-pass fluid flow during crustal thickening, and by conduction only after crustal thickening. Parameters: *f*<sub>end</sub>=1.6; *D*=10 km; *A*<sub>0</sub>=2.5 W/m<sup>3</sup>; ρ<sub>crust</sub>=2700 kg/m<sup>3</sup>; boundary conditions of 0 and 1350°C (*T*<sub>max</sub>); no erosion or tectonic denudation after thickening; and an initial lithospheric thickness, *L*, of 100 km.

**A:**

- *ε*<sub>experiments 1, 2, 3</sub> = 10<sup>−15</sup> 3 s<sup>−1</sup> (*W* = 2.5 10<sup>−12</sup> m<sup>−1</sup> s<sup>−1</sup>);
- *ε*<sub>experiments 4, 5, 6</sub> = 10<sup>−14</sup> 82 s<sup>−1</sup> (*W* = 7.6 10<sup>−11</sup> m<sup>−1</sup> s<sup>−1</sup>);
- *ε*<sub>experiments 7</sub> = 10<sup>−14</sup> 3 s<sup>−1</sup> (*W* = 2.5 10<sup>−11</sup> m<sup>−1</sup> s<sup>−1</sup>). In experiment 1, 4 and 7, *t*<sub>det</sub> = 0 Ma.; in experiment 2, *t*<sub>det</sub> = 7.5 Ma.; in experiment 3, *t*<sub>det</sub> = 15 Ma.; in experiment 5, *t*<sub>det</sub> = 2.5 Ma.; in experiment 6, *t*<sub>det</sub> = 5 Ma.. Additional parameters: *K* = 3 W/mK; *κ* = 1.2 10<sup>−6</sup> m<sup>2</sup> s<sup>−1</sup>. Initial crustal thickness: 25 km. The stable geotherm includes a 4 km thick granite with a radiogenic heat production of 4.1 μW/m<sup>3</sup> between the 18 and 22 km depth levels. The latter heat source distribution is thought to represent the Proterozoic situation of the Mount Isa Inlier.

**B:**

Four additional experiment results (curves 5 and 6 are the same as respectively curves 3 and 1 in Fig. 4.9A).

- *ε*<sub>experiments 1, 3</sub> = 10<sup>−14</sup> 3 s<sup>−1</sup> (*W* = 2.5 10<sup>−11</sup> m<sup>−1</sup> s<sup>−1</sup>);
- *ε*<sub>experiments 2, 4, 5, 6</sub> = 10<sup>−15</sup> 3 s<sup>−1</sup> (*W* = 2.5 10<sup>−12</sup> m<sup>−1</sup> s<sup>−1</sup>). *ε*<sub>experiments 1, 2, 3, 4</sub> = 10<sup>−6</sup> m<sup>2</sup> s<sup>−1</sup>. Experiments 1 and 2 (curves 1 & 2) simulated instantaneous underplating of a 25 km thick and 1200°C hot slab at the base of the crust (at 35 km) at the onset of thickening. During thickening this slab cools by conduction only. Experiment 3, 4, 5 and 6 simulated crustal thickening (initial crustal thickness in experiment 3 & 4: 35 km; in exp. 5 & 6: 25 km) accompanied by convective thinning of the mantle lithosphere. In experiment 1, 2, 3 and 4, the steady-state conductive geotherm is defined by

\[
T_x = \frac{A_0 D^2}{K} \left(1 - e^{-xD/K} \right) + \frac{T_{max} A_0 D^2}{L} \frac{K}{x}, \quad x \text{ in km}
\]

(with *K* = 2.5 W/Km). In experiment 3, 4 and 6, *t*<sub>det</sub> = 0 Ma., in experiment 5, *t*<sub>det</sub> = 15 Ma.. Steady-state temperatures for the thickened crust are marked by blocks (exp. 1, 2, 3, 4) and triangles (exp. 5, 6). Maximum temperatures reached after thickening are marked by the experiment number. Experiments like number 6 (number 1 in Fig. 4.9A), in which fast convective thinning of the entire mantle lithosphere accompanies relatively slow crustal thickening, yield *P*-*T*-*t* paths similar to the ones deduced from the Early to Middle Proterozoic fold belts of Australia (e.g. trajectory "RH", which has been petrologically deduced by Reinhardt & Hamilton, in press).
Numerical modelling

The lithosphere was simulated by a sudden shift of the lower boundary condition (T=1250°C to 1350°C) to the base of the (deforming) crust. Thus, a situation is simulated in which the complete mantle lithosphere is detached from the crust (thermally, this model is identical to crust-mantle delamination). At the end of convection (assumed here to coincide with the end of crustal thickening), the lower boundary condition (T=1250°C to 1350°C) was abruptly shifted back to the base of the pre-deformation lithosphere. Cooling is purely isobaric if no erosion or tectonic denudation takes place. Those experiments characterized by a relatively fast convective thinning, low strain rates, long convection times at the base of the crust, and slow or no erosion yield anti-clockwise P-T-t paths. Figure 4.9A illustrates this, showing seven P-T-t paths, all from experiments with a 25 km thick initial crust, a finite crustal thickening of 60%, and no erosion at the end of crustal thickening. The initial time of crust-mantle detachment in these seven experiments varied, but once detached, the temperature at the base of the crust was held constant till the end of deformation (hot mode).

4.4 IMPLICATIONS OF MODELS FOR THE MOUNT ISA INLIER

4.4.1 lithospheric thinning

4.4.1.1 asymmetric extension

Following and extending McKenzie's (1978) arguments for thermal relaxation after a pure-shear extensional event, the three documented rifting events of the Mount Isa Inlier, of 1780, 1750 and 1680 Ma. respectively, are unlikely to be related to the low-P facies metamorphism at 1600-1550 Ma., primarily because the time gap between the extensional events and metamorphism is too wide. The same thermo-mechanical arguments refute the suggested (Hobbs et al., 1984) relation between the regional, prograde, low-P facies metamorphism at 1600 Ma. and ensialic rifting at approximately 1700 Ma. (R.W. Page, pers. comm., 1988) in the Broken Hill Block, Australia. In contrast, since peak metamorphic conditions were reached at the end of F₁ (early deformation; 10 km scale recumbent folds), high-P facies metamorphism would be expected.

The fact that crustal thinning proves to be in many cases an asymmetric process, resulting in two different, but complementary plate halves, cannot overcome this problem. Thermal relaxation times are of the same order as the pure-shear model. The main problem with any model incorporating late asymmetrical extension is the absence of young (post-1670 Ma.) rift sediments and late extensional structures. Although the asymmetric extension models explain lateral offsets between areas of upper crustal
extension and areas affected by low-P facies metamorphism, on the scale of the complete inlier (400 x 200 km), normal faulting would have to be developed at least in some areas; even more so, since transfer faults would compartmentalize the extensional terrane, so that upper and lower plates are juxtaposed along strike (Gibbs, 1984; Bosworth, 1985; Etheridge, 1987; Lister et al., in press).

### 4.4.1.2 magmatic events

Magmatism results from a primary perturbation of a depth-temperature profile and is thus secondary in time. More importantly here, its effect is also secondary on this perturbation. There is no question that during crustal thickening granitoid lower crust may melt, and that during (asymmetric) extension deeper-seated pressure-release melting may occur. However, whatever the mode of magmatism, the pulse of magmatic heat is short-lived and T/P ratios will only temporarily be raised. In the Mount Isa Inlier, there is no evidence for syn- or immediately pre-metamorphic granites. In the Broken Hill Province, syn-tectonic granites are present (Hobbs et al., 1985), as in most low-P terranes (Kerrison, written comm., 1989).

Still, the thermal effect of mid-crustal, pre-metamorphic granites can be strong and will be long-lived, as most of them are enriched in HPE. Table 4.2 gives concentrations of the HPE in the main pre-metamorphic granites of the Mount Isa Inlier.

<table>
<thead>
<tr>
<th>Table 4.2</th>
<th>Concentration of heat producing elements (HPE) in the three largest pre-metamorphic batholiths of the Mount Isa Inlier, as compared to the worldwide average concentration in granites. The percentages under the batholith names represent the surface coverage of the respective granites, so that a weighed average heat production can be calculated. Data supplied by L.A.I. Wyborn.</th>
</tr>
</thead>
<tbody>
<tr>
<td>worldwide average</td>
<td>Sybella (3.14%)</td>
</tr>
<tr>
<td>low Ca</td>
<td>high Ca</td>
</tr>
<tr>
<td>3</td>
<td>8</td>
</tr>
</tbody>
</table>

The concentration of HPE in Mount Isa's granites is anomalously high; it yields an average radiogenic heat production of 4.1\mu W/m³. This is 52% higher than the worldwide average (2.68\mu W/m³ according to Turcotte & Schubert, 1982). Figure 4.1 shows the effect of a 4 km thick granitic slab between 18 and 22 km with a heat production of 4.1\mu W/m³ on the steady-state conductive geotherm. It will raise the temperature in the 3-5 kbar interval by 50-75°C, which is not enough to transect the andalusite and sillimanite stability fields. The temperature change (T last /T last-1) decreases
linearly with distance to the granite (Fig. 4.7). The HPE-enrichment within the Sybella Batholith e.g. can explain the (regional) low-P facies metamorphism within and closely around it.

Pre-metamorphic HPE-enriched granites are widespread elsewhere in the northern Australian Early to Middle Proterozoic fold belts (Wyborn et al., 1987). The presence of such granites alone can explain steep average geothermal gradients, but not the anti-clockwise P-T-t path. Also, no major pre-peak-metamorphic granites are within 20 km (surface-) distance to the Soldiers Cap Group (SCG), whereas it too is affected by the low-P facies metamorphism (Jaques et al., 1982).

If HPE-enriched granites were responsible for the low-P facies metamorphism, an anomalously high thermal gradient could be expected to affect the inlier also at present. Hyndman & Sass (1966) and Cull (1982) calculated an average geothermal gradient in the upper crust of 19.8°C/km, and heat flows of respectively 81.9 and 75.3 mW/m². Even when compared to the Australian background heat flow (which is the highest of all continents), this is high: a worldwide continental average is 56.5 mW/m² (=1.35 HFU; Turcotte & Schubert, 1982). However, thermal gradients of 20°C/km are still far too low to cause andalusite-sillimanite metamorphism at depth.

4.4.1.3 Crust-mantle detachment

Figure 4.5F shows that sillimanite grade conditions can be reached when the crust is abruptly "underplated" by hot asthenospheric material, which, after upwelling, cools by conduction. It simulates in simplified form the extension-triggered delamination models for Proterozoic fold belts, proposed by Kröner (1983) and Etheridge et al. (1987b). Such models, however, cannot be applied to the Mount Isa Inlier for the same reasons that models incorporating only extension cannot be applied: the time gap between extension/delamination and low-P facies metamorphism is too wide. Hot mode delamination, associated with the documented extensional events, would mitigate this objection. However, there is little evidence for anomalously steep geothermal gradients in the 120 Ma. time gap between extension and metamorphism: there is no record of major crustal magmatism in this time span, nor is Dc1 generally characterized by low-P facies assemblages. Similarly, in the Willyama Block (Fig. 1.1), metamorphic grades increase during crustal thickening (Hobbs et al., 1984). Any causal relation between the documented extensional events in the inlier and the at least 120 Ma. younger low-P facies metamorphism is therefore invalid. In fact, regional, prograde metamorphism (+∂T/∂t) in general cannot be explained by any positive pre-metamorphic thermal perturbation (extension, magmatism, crust-mantle delamination), as this would only lead to a decrease
of the perturbed temperatures (-∂T/∂t). A variation, favoured here, is a coupling between crust-mantle detachment and the compressional events.

4.4.2 crusted thickening and concomitant convective thinning of the mantle lithosphere

The main attraction of this model is that convective thinning of the mantle lithosphere is synchronous with crustal thickening (and thus the period of low-P facies metamorphism), and heat flow at the base of the crust may remain high during the entire phase of crustal thickening. Thus, unlike the short-lived thermal perturbation brought about by extension and magmatism, isotherms may stay elevated for a considerable time. The ease and speed of convective lithospheric thinning depend on the viscosity of the upper asthenosphere. As viscosity is very sensitive to temperature, slightly steeper geothermal gradients in the Precambrian might explain the prevalence of Precambrian low-P facies metamorphism. On the other hand, there appear to be two problems with this model. Neither is, however, insuperable.

First, it is unlikely that this model can lead to crystallization of sillimanite during the early stages of crustal thickening, since the delamination (imagmatic events) would have to predate most of the crustal thickening and strain rates would thus have to be very low. In the Mount Isa Inlier, however, peak metamorphic low-P facies assemblages are in most cases coeval with the early stages of \( D_{c2} \). If thinning of the mantle lithosphere was solely the result of \( D_{c2} \), strain rates must have been unrealistically low (<\( 10^{-15} \) s\(^{-1} \)) and detachment of the upper crust relatively fast. It is more likely that a major thickening event preceded \( D_{c2} \), i.e. \( D_{c1} \) was not just of local importance, but triggered thinning of the mantle lithosphere. \( D_{c1} \) thrusting has now been documented in the western part of the inlier (Bell, 1983), in the central part (Loosveld & Schreurs, Appendix 1) and in the eastern part (Loosveld, Chapters 2 & 3). It is noteworthy in this respect that the early crustal thickening, \( D_{c1} \), is characterized by pre-dominantly brittle deformation, whereas ongoing crustal thickening, \( D_{c2} \), is marked by dramatically higher temperatures and predominantly ductile deformation. This is regarded as typical for the Early to Middle Proterozoic fold belts of northern Australia (Etheridge et al., 1987b), although \( D_{1} \) in the Broken Hill Province is ductile (Hobbs et al., 1987).

The second apparent problem lies in the post-tectonic history. Generally, crustal thickening of a 35 km thick crust must lead to syn- to post-orogenic uplift, either controlled by erosion or by late-compressional extension. In the Mount Isa Inlier, however, petrological constraints indicate compression followed by essentially isobaric cooling (Reinhardt & Hamilton, in press). A possible explanation for this failing uplift may be that the crust was abnormally thin at the onset of crustal thickening. Thickening of this thin crust may then have resulted in a thick, but not exceptionally thick and
therefore gravitationally metastable crust. Refraction seismics (Finlayson, 1982) and gravity data (Wellman, 1976; Shirley, 1979; Dooley, 1980) support this possibility, as they indicate that the crust under the Mount Isa Inlier is at the moment 40-45 km thick. Adding the eroded package of 10-15 km gives a ~55 km thick crust at the end of the crustal thickening events. Assuming a finite thickening factor of 2 for these events, and no syn-Dc2 underplating, the crust would have been only ~28 km thick prior to thickening. If underplating into the lower crust accompanied Dc2, the 28 km is a maximum (however, if extension factors were smaller than 2, only little melt would be generated, according to McKenzie & Bickle, in press).

A more general explanation, however, exists: as time constants for erosion (λ: 60-300 Ma.; England & Richardson, 1977) are generally much greater than thermal time constants (heat transfer by conduction only: L2/π2k, in the order of tens of Ma.), the immediately-post-thickening segment of the P-T-t path will generally be characterized by slightly decompressive cooling. (Note, however, that upwelling of hot asthenospheric material, as proposed here, would tend to increase the uplift triggered by crustal thickening, and in fact was originally postulated by Bird, 1978, 1979, to explain intraplate heating and uplift).

The 1500 Ma. old granites, the Williams and Naraku Batholiths, clearly postdate Dc2. The main, coarse-grained phase of the Naraku Granite gives a (Sm/Nd) TDM source age of 1637 Ma.; the TDM source age of the Williams Batholith ranges from either 1530 to 1620 Ma., or, and more probably, from 1630 to 1720 Ma. (Wyborn et al., in press). (The LREE-depleted, near-mantle εNd values (+2.0 to +3.5) for the post-tectonic granites also imply a relatively young source emplacement.) Thus, the source ages for the main phases of the Williams and Naraku Batholiths are close to the deformation ages for Dc1 and (less so) Dc2, respectively 1610±12 and 1544±13 Ma.. Although the chronological database is still very small (and the source ages very model-dependent), the crustal thickening period(s) may have been accompanied by major mantle events during which huge amounts of source rocks were created in the lower crust or just under the crust (Wyborn et al., in press). A concurrence of crustal thickening and such mantle events would strongly point towards Houseman et al.'s (1981) model for convective instability of the mantle lithosphere during compressive tectonics.
4.5 CONCLUSIONS

General conclusions are:

(1) Cooled granitoids, enriched in U, Th and K, may substantially contribute to the local heat production.

(2) Prograde metamorphism (+ΔT) cannot be attributed to any pre-metamorphic heating event. Such pre-metamorphic events can explain low-P facies metamorphism, but not the prograde character of it.

(3) Compressive (+ΔP) low-P facies metamorphism cannot be attributed to crustal thickening alone, as this gives rise to clockwise P-T-t paths with possible low-P facies metamorphism during post-thickening decompression.

(4) After elimination of other options, like single-pass fluid flow during Dc2, crustal thickening and associated convective thinning of the mantle lithosphere remains as the only possible explanation for anti-clockwise P-T-t paths. Convective thinning of the mantle lithosphere has to be fast with respect to the progressive crustal thickening. Temperatures at the base of the crust have to remain high until the end of thickening.

Conclusions specifically concerning the Mount Isa Inlier (and probably other Early to Middle Proterozoic Australian fold belts) are:

(1) The P-T-t paths for the Early to Middle Proterozoic fold belts of Australia are markedly anti-clockwise, made up of a segment of prograde, low-P facies metamorphism (compressive heating; +ΔT, +ΔP), followed by a segment of essentially isobaric cooling. In the Mount Isa Inlier, the low-P facies metamorphism is contemporaneous with Dc2, a phase of pervasive crustal thickening. Extension and pre-metamorphic magmatism predate this major crustal thickening event by at least 120 Ma.

(2) Granitoids, enriched in U, Th and K, substantially contribute to the local heat production.

(3) In the Mount Isa Inlier, convective thinning of the mantle lithosphere might have been triggered by Dc1, the early thrusting event.
CHAPTER 5

SYNTHESIS

LOOSVELD
The Mount Isa Inlier has a protracted Early to Middle Proterozoic history of mafic underplating, basin development, felsic and mafic magmatism, crustal thickening and metamorphism. The oldest traceable event comes from the Ewen and Kalkadoon Batholiths, the Leichhardt Metamorphics, the Argylla Formation, the Wonga Batholiths and the Lunch Creek Gabbro, which all have a common Sm-Nd signature of 2140-2300 Ma. (Wyborn et al., in press). These dates are interpreted to represent a major phase of underplating at the base of the crust. Underplating, however, was probably episodic, as other TDM Sm-Nd dates are 1790 (Sybella Batholith) and 1530-1720 Ma. (Williams Batholith). The tectono-thermal triggering mechanisms for these underplating events are contentious.

The evolution of the inlier from 1860 Ma. onwards has recently been interpreted as continuously ensialic, with pre-1860 Ma. old basement inliers overlain by three Mid-Proterozoic cover sequences. All three cover sequences have been interpreted as rift (and sag) deposits, mainly on the grounds of their lithological contents, but also on some evidence for extensional structures. The tectono-stratigraphic framework of the eastern part of the inlier was more obscure; geochronological dates are scarce. In the field study of this project, I have set out to obtain a better understanding of the basin-forming events in the central eastern part by unravelling the basin-modifying structures. In this central eastern area, the basement and oldest cover sequence are not exposed. Beardsmore et al. (in press) argue that the Soldiers Cap Group (SCG) might represent the oldest cover sequence, but evidence for this correlation is weak, and the group most likely belongs to the second cover sequence (D2). Considering the problems which surround the correlation of the SCG, geochronological dating of the group is crucial.

In this study, the central eastern part of the inlier is subdivided into three tectono-stratigraphic domains, which differ considerably from each other lithologically, but which have essentially undergone the same deformational events. Orientation of dominant structures, however, varies greatly, and correlation of structures on the grounds of orientation leads to erroneous results. All meta-sediments in the study area, except for possibly some remobilised calc-silicate breccias, have been attributed in this study to the second cover sequence, which was deposited between 1780 and 1740 Ma.

The second cover sequence (including rift packages: the Argylla Formation, the lower Malbon group, the lower SCG, and sag packages: the upper Malbon Group, the upper SCG and the lower Mary Kathleen Group) results from intra-continental extension at ≈1780 Ma. The upper Mary Kathleen Group, containing the Tommy Creek Block and the Roxmere Quartzite, is deposited during a 1750±10 Ma. extensional event, D2.
Evidence for cover sequence 3 in the central and eastern part of the inlier (Mount Albert Group) should be critically reviewed. I question the widely-accepted view that the Deighton and Roxmere Quartzites belong to the third cover sequence (respectively in appendix 1 and chapter 2): the Deighton Quartzite (and White Blow Formation), although concordantly overlying what has been defined Mary Kathleen Group (Derrick et al., 1977a), may well be a lateral facies equivalent of the Mary Kathleen Group, whereas the Roxmere Quartzite is conformably under- and overlain by calcisilicate rocks of presumably the upper Mary Kathleen Group, and thus is probably part of this group too.

Over the entire inlier, basement and cover sequences alike were affected by a penetrative phase of crustal thickening, $D_{c2}$, characterized by upright, non-cylindrical, N-S to NE-SW trending folds and vertical extension lineations, and accompanied by regional, low-P facies metamorphism. An age of 1550 Ma. is reasonable well constrained for this event. Locally, the development of thrusts and fold nappes preceded $D_{c2}$, giving rise to a rather erratic pattern of movement vectors. If these structures can be attributed to the same event, $D_{c1}$, geometrically necessary tear faults should have been developed between the domains of different movement vectors. $D_{c1}$ should be younger than 1670 Ma., the age of the Mount Isa Group. Zones of intense $S_1$ development in the Syabella Granite yielded Rb-Sr whole rock dates of 1610±13 Ma. (Page & Bell, 1986). A conjugate set of NNW- and NNE-trending strike-slip faults, E over W reverse faults, crenulations and kink bands of various orientations, and N-S reverse reactivation of early extensional faults all postdate $D_{c2}$. Most of these structures affect the ≈1500 Ma. old Williams and Naraku Batholiths, and are possibly associated with a widespread resetting of K-Ar ages around 1500-1450 Ma.. $D_{c3}$ in the Syabella Granite was dated by the Rb-Sr whole rock technique at 1510±13 Ma. by Page & Bell (1986).

The eastern part of the inlier, although typified by a different litho-stratigraphical and magmatic architecture and by different orientations of the main structures, has undergone the same three contractional events. $D_{c1}$ resulted in the Snake Creek fold nappe and the development of E-W trending, upright to steeply S-dipping, isoclinal folds, the overturning of some Roxmere Quartzite masses, and localized development of slaty cleavages in the Marimo-Staveley Block and Mitakoodi Anticlinorium. $D_{c2}$ is the penetrative deformation event, resulting from E-W compression. The Mitakoodi Anticlinorium, most N-S trending folds in the Marimo-Staveley Block, the Weatherley Creek Syncline, the Middle Creek Anticline and the Pumpkin Gully Syncline are $D_{c2}$ structures. This phase of deformation is accompanied by low-P facies metamorphism. $D_{c3}$ is manifested by a variety of non-penetrative structures such as NNW-trending reverse faults (the Cloncurry "Overthrust") and WNW-trending folds.
This structural framework sheds light on litho-stratigraphic relations. It is concluded that the Marimo-Staveley Block lithologies (Mary Kathleen Group) may well directly overlie the Malbon Group and SCG. Previous interpretations of a continental margin based on an eastward-deepening basin (Argylla Formation and Malbon Group: sub-aerial; Mary Kathleen Group: shelf; SCG: deepwater turbidites) are therefore not justified. The Mount Isa Inlier does not represent the eastern margin of a Mid-Proterozoic continental margin, because of (a) the absence of rocks of oceanic crust affinity in the inlier, (b) the paucity, or absence even, of igneous suites of continental arc affinity, (c) the consistent continental tholeiitic affinity of mafic rocks both west and east in the inlier, (d) the symmetrical thinning of the crust east and west of the inlier, rather than an asymmetric thinning at its eastern side, (e) the occurrence of equally old continental rocks east of the inlier, and (f) the absence of high-P facies metamorphic assemblages. The Soldiers Cap Group, the easternmost exposed group of the inlier, can be interpreted as being deposited in a fairly elongate, deepening rift, synchronous with other rock units of the second cover sequence (1780 Ma.). Extension, however, did not at any stage reach factors at which oceanic crust was generated.

The second part of the thesis addressed the problem of the rather enigmatic synchronicity of crustal thickening, Dc2, and the low-P facies metamorphism. Two fundamental explanations have been tested, (a) an as yet not documented lithospheric extension immediately predating Dc2, and (b) crustal thickening associated with the basal erosion (=thinning) of the mantle lithosphere.

Extension directly followed by Dc2 crustal thickening might explain why the crustal thickening is not followed by exhumation, either by erosion or by tectonic denudation. There are two pressing arguments, however, against such an extensional model. Firstly, no extensional structures have been documented in the Mount Isa Inlier (nor in the Willyama Province) that could predate Dc2 by less than the thermal time constant of thermal readjustment of a thinned (or even a normal) lithosphere. At Dc2 times, the geotherm would have recuperated from any major disturbances due to the documented phases of lithospheric extension. Secondly, it was shown that, based on a reasonable, actualistic steady-state, conductive geotherm, extension factors in the order of 3 are needed to reach andalusite-sillimanite conditions during extension. Thus, one can expect andalusite-sillimanite assemblages only in extremely thinned parts of the crust, as e.g. passive continental margins, or in areas with anomalously steep pre-extensional geothermal gradients. Additionally, metamorphism in the inlier during crustal thickening is prograde (+AT), and not, as would be expected after a more than a few Ma. old extensional event, retrograde (-ΔT).
During crustal thickening sensu stricto, temperatures should theoretically increase adiabatically (in the case of instantaneous thickening), or increase slightly faster than that depending (inversely proportional) on the thickening rate. In the Early to Middle Proterozoic fold belts of Australia, however, the increase in temperature during crustal thickening is considerably larger (initially even increasing T/P ratios) than that which could be explained by low strain rates, and one has to appeal to an extra heat source. The direct relation between the low-P facies metamorphism and the crustal thickening implies the relation to be causal. Simple numerical simulations of the "crustal thickening/convective lithospheric thinning" model, using reasonable parameters, yield P-T-t paths which may transect the andalusite-sillimanite stability field in a prograde (+ΔT, +ΔP) fashion. After thickening, when high-level mantle convection (at the base of the crust?) stops, the mantle cools conductively. As time constants for erosion are generally much greater than thermal time constants, this cooling phase will be isobaric to slightly decompressive. The numerical experiments reveal that if lithospheric thinning is fast relative to crustal thickening, and if temperatures at the base of the crust remain high until the end of deformation, the resulting P-T-t paths will be anti-clockwise and, depending on the original steady-state geotherm, similar to the ones petrologically deduced for the Proterozoic fold belts of Australia.

Timing relations between low-P facies metamorphism and crustal thickening in other "Hercyno-type" fold belts should be reviewed. P-T-t trajectories may well contain the key to the tectono-thermal history of these orogenies too.
ADDENDUM REFERENCES


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APPENDICES
APPENDIX 1

DISCOVERY OF THRUST KLIPPE, NORTHWEST OF MARY KATHLEEN, MT. ISA INLIER, AUSTRALIA

LOOSVELD
SCHREURS

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The "crustal thickening/convective lithospheric thinning" model, as numerically modelled by Houseman et al. (1981) and used here as an explanation for the prograde low-P facies metamorphism that accompanied Dc2 in the Mount Isa Inlier (Chapters 3 & 4; Appendix 8), is characterized by anomalously high T/P ratios during the later increments of, and also immediately after, the crustal thickening event which triggers the "erosion" (thinning) of the mantle lithosphere. (In the experiment series of Houseman et al. (1981), the peak of kinetic energy, associated with the main lithospheric thinning, postdates the instantaneous lithospheric thickening by 4-88 Ma.) In the Mount Isa Inlier, however, andalusite/sillimanite blastesis is in places synchronous with the early phases of Dc2 (e.g. Fig. 3.4). Therefore, for the "crustal thickening/convective lithospheric thinning" model to hold, it is imperative that Dc2 was preceded by another phase of crustal thickening, and it becomes crucial to establish the nature of the early deformation. Two contrasting theories exist. The first one, by Bell (1983), Loosveld & Schreurs (Appendix 1) and Loosveld (Chapters 2 & 3) states that a major thrusting event, Dc1, preceded Dc2 by approximately 65 Ma. (Page & Bell, 1986). This model has been criticized and modified by Stewart (BMR, 1988), who interpreted the early structures as resulting from extension and only minor, later (but still pre-Dc2) thrusting. Both the Loosveld & Schreurs (1987) and the Stewart (BMR, 1988) studies were carried out in the central Mary Kathleen Sheet area. Now follows the Loosveld & Schreurs study, the "Discussion" by Stewart (BMR, 1988), and the "Reply" by Loosveld.
Discovery of Thrust klippen, northwest of Mary Kathleen, 
Mt Isa Inlier, Australia

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In the northeastern part of the Mary Kathleen 1:100 000 sheet, a thick pile of Early to Middle Proterozoic metasediments, metavolcanics and intrusive rocks were deformed by an early thrusting event and later by east–west shortening. D1 produced a thrust complex, containing imbricate stacks of thrust sheets and a sequence of foliations, changing in orientation and style: S1A is a phyllonitic fabric on the major thrusts; S1B is the axial-plane foliation to west-verging folds in the immediate vicinity of the major thrusts; and S1C is the axial-plane foliation of northeast–southwest upright folds, locally developed within the thrust sheets, caused by shortening parallel to the transport direction. Two thrust systems have been studied in detail. In the first, in the Deighton area, the geometry of the imbricates, the extension lineations on S1A and the orientation of D1C folds all point towards a westward or northwestward movement direction. In contrast, the West Leichhardt imbricates may have moved southward. After thrusting, both systems underwent east-west shortening during D2, causing the alignment of elongate doubly plunging klippen, separated from each other by underlying steepened imbricate stacks. These results require a revision of the significance of the current lithostratigraphic column: the possibility that a major tectonic repetition rather than a sedimentary repetition occurs in the cover sequences cannot be excluded at present.

Key words: Mt Isa, Proterozoic mobile belts, Queensland, structural geology, thrusting.

INTRODUCTION

The central Mt Isa Inlier consists of Early to Middle Proterozoic, medium grade metasediments, metavolcanics and intrusive rocks, which have undergone an extensive east–west ductile-shortening event, associated with vertical extension. This deformation phase, D2, is penetrative and recognizable over the entire central Mt Isa Inlier. D3 is less penetrative and produced an apparent conjugate strike–slip fault system, of which the Wonga Fault and the Trey Bit Fault are examples (Fig. 1). Because of a lack of detailed knowledge of the stratigraphic sequence and extensive overprinting by the D2 and D3 deformations, all pre-D2 structures are difficult to recognize and interpret. Clearly these early deformations are dominated by low angle faults, which for different areas have been variously interpreted as: due to horizontal shortening (Bell 1983); and as extensional because of the existence of listric normal faults (Dunnet 1976), and a local brittle extensional detachment (Passchier 1986). This paper documents areas with low angle faults, one of which is of particular interest because it is in the same tectonostratigraphic zone as Passchier's extensional terrane.

GEOLOGICAL SETTING

All rocks in the inlier are of Early to Middle Proterozoic age and occur in a structurally complex setting. Three broad structural divisions outlined by Carter et al (1961) consist of eastern and western successions, separated by the Kalkadoon–Leichhardt Block. An upgraded subdivision by Blake (1987) consists of a Western and an Eastern Fold Belt, separated by the Kalkadoon–Leichhardt Belt. Considerable dispute still exists as to what extent the successions of the Eastern and the Western Fold Belts can be correlated (compare Blake 1980 with Derrick et al 1977). In this paper the stratigraphic nomenclature established by Derrick et al (1977) is followed.

The only rocks attributed to the Western Succession studied in this work are represented by a narrow north–south belt of conglomerate, quartzite and siltstone named the 'West Leichhardt Klippe' (Fig. 1). It was mapped by Derrick et al (1977) as the Surprise Creek Beds.
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Fig. 1 Structural map of the northeastern part of the Mary Kathleen 1:100 000 Sheet area, plus a schematic representation of the lithological succession dealt with in this paper.
although the same rocks are placed in the Quilalar Formation on the adjacent 1:100 000 Prospector Sheet to the north.

The Kalkadoon–Leichhardt and Blockade blocks (Fig. 1) consist mainly of the Kalkadoon Granite and the Tewinga Group. The Kalkadoon Granite, a large I-type batholith of mainly granitic to granodioritic composition, covers most of the central part of the Mary Kathleen Sheet area and has been dated (U-Pb Zr) at 1856 Ma (Wyborn & Page 1983). The Tewinga Group comprises three formations, from oldest to youngest as follows:

(i) The Leichhardt Volcanics consist of greyish porphyritic metarhyolite with lesser amounts of rhyodacite and dacite. These acid metavolcanics yield a U-Pb Zr age of 1867 ± 5 Ma (Page 1983) and are intruded by the Kalkadoon Granite;
(ii) The Magna Lynn Metabasalts, a succession of massive and amygdaloidal metabasalt; and
(iii) The Argylla Formation, dated at 1777 ± 7 Ma (Page 1978), which consists of acid metavolcanics and minor metasediments.

The Eastern Succession covers the largest portion of the area and consists of a meta-sedimentary sequence, divided in previous work (Derrick et al. 1977) into the Mary Kathleen Group and the younger Mt Albert Group. Both groups consist of two formations, the lower being quartzitic and the upper pelitic/calcareous. The latter, certainly in the Mary Kathleen Group, has a much wider distribution than the lower quartzitic formations. Such large-scale lithological repetitions could be interpreted as a reflection of repeated rift and sag phases during intracontinental extension but in this paper we will suggest that the Mary Kathleen Group is a lateral facies equivalent of the Mt Albert Group, the latter having been thrust over the former.

The Mary Kathleen Group

The Mary Kathleen Group in the area consists of the Ballara Quartzite and the Corella Formation. The former, usually a pure quartzite which forms rugged ridges, conformably overlies the Argylla Formation; locally the contact is marked by a basal conglomerate containing clasts of the underlying unit. Abundant sedimentary structures such as (trough) cross-bedding and ripple marks are recognizable despite metamorphism and deformation. The lower Ballara Quartzite and conglomerate in the lower part of the Ballara Quartzite represent ‘deltaic or coarse fluvialite deposits, or nearshore cliff or beach deposits’ and sediments in the upper part of the formation a ‘shallow shelf subjected to moderate current action’ (Derrick et al. 1977). The Corella Formation conformably overlies the Ballara Quartzite and contains thin-bedded metasediments of mostly calcareous/pelitic or psammitic composition and minor mafic metavolcanics.

The Mt Albert Group

The Mt Albert Group in the Mary Kathleen Sheet area was divided by Derrick et al. (1977) into the Deighton Quartzite and the White Blow Formation. The Deighton Quartzite consists mainly of cross-bedded quartzite and minor metasiltstone, and according to Derrick et al. (1977) was originally a fluvialite deposit. However, symmetrical ripples, indicating wave interaction and therefore near-shore environments, have been observed at the southern tip of the southernmost mass of the Deighton Quartzite (GR 858969; numbers given in this way will hereafter refer to the Mary Kathleen 1:100 000 Sheet; see locality at point A in Fig. 1). The White Blow Formation consists of inter-bedded phyllite, feldspathic quartzite, black slate and minor laminated calc-silicate rocks.

METAMORPHIC AND DEFORMATIONAL HISTORY

Most exposed rocks in the inlier have undergone greenschist to lower amphibolite facies metamorphism as indicated by quartz–biotite–muscovite–garnet ± andalusite assemblages in the pelitic metasediments and quartz–biotite–muscovite–scapolite–calcite assemblages in the calcareous metasediments. Low pressure amphibolite facies parageneses in the Deighton area (Fig. 1) have been described by Derrick et al. (1977) and for an area 60 km to the south by Passchier (1986).

Three main structural events have been recognized. We treat them from youngest, $D_3$, to
Fig. 2 A D$_2$ shearzone in beds of the Corella Formation (GR 720077; point C in Fig. 1). The folds are typical D$_2$ folds.

Fig. 3 Basal, polymict conglomerate at the contact between the Argylla Formation and the Ballara Quartzite (GR 721077; point D in Fig. 1). The marker pen is parallel to L$_3$. Prolate pebbles are rotated towards this vertical direction.

oldest, D$_1$, because complexity increases in this order and the emphasis in this paper is on D$_1$.

**D$_3$: late deformation associated with strike-slip faulting**

D$_3$ is an assemblage of post D$_2$ events (Lister et al. 1986), uncorrelated because of their non-penetrative nature, represented by conjugate sets of shear zones and faults of varying size which transect the area. The dominant trend in the north is 030°, examples of which are the Wonga Fault and the Trey Bit Fault (Fig. 1), whereas in the south there are both 330° and 030° trending shear zones and faults. Sense of displacement is generally extremely difficult to establish, although a dextral strike-slip component seems to dominate along the 030° trending faults. Minor concentrations of malachite and azurite are present near D$_3$ shear and fault zones (e.g. GR 837052; point B in Fig. 1).

Open, upright, east-trending folds, observed only around the West Leichhardt Klippe, are also grouped in D$_3$ since they overprint chloritic D$_2$ shear zones within the Kalkadoon Granite just east of this klippe. Kinks with various orientations are the youngest observed structures.

**D$_4$: penetrative deformation associated with east–west shortening**

This is a major deformational event, involving considerable east–west shortening which produced approximately north–trending, upright folds. A penetrative axial-plane foliation is commonly developed and varies from a spaced
cleavage to a slaty cleavage. \( D_2 \) folds are often accompanied by steep shear zones (Fig. 2), especially on the limbs of kilometre-scale \( D_2 \) folds. Fold axes have rotated from horizontal in the hinge of these large folds to vertical in the shear zones on the limbs. The extension lineation in \( S_2 \), \( L_1 \) is defined by elongated quartz aggregates and is vertical or steeply plunging. Sedimentary imbricated pebbles in basal conglomerates are elongated in the \( XY \) plane (e.g. GR 721077; point D in Fig. 1). Deformed pebbles exhibiting oblate and prolate ellipsoidal forms are shown in Fig. 3: the long axis, \( X \), is vertical, whilst the \( XY \) plane is north-south vertical and coincides with the axial planes of the \( D_2 \) folds.

Due to variable orientations of the older \( D_1 \) axes and \( D_1 \) axial planes, a wide range of fold interference patterns (Ramsay 1967) can occur where \( D_2 \) overprints \( D_1 \), even at outcrop scale. However, many of the striking, map scale, doubly plunging synclines (e.g. the Deighton and Robin Synclines) can be attributed to \( D_2 \) alone: vertical extension during \( D_2 \) varied from place to place, producing highly non-cylindrical folds.

Previous workers in adjacent areas (Bell 1983; N. Oliver pers. comm.; P. Pearson pers. comm.) described upright, 340° trending folds, which they attributed to \( D_3 \). These fold orientations are also abundant in the Mt Devine Imbricate Stack (Fig. 1), north and northwest of the Deighton Syncline. However, because there are no convincing overprinting relations between the north-trending \( D_2 \) folds and the 340° trending folds and both display the same fold style, we prefer to group the 340° trending folds with the \( D_2 \) folds.

**\( D_1 \): early deformation associated with low angle faulting**

The oldest and by far the most complex deformation was produced by early low angle faulting, associated with a great variety of folds and ductile shear zones. These low angle faults (i.e. low angle to bedding but generally with steep dips) have been interpreted as extensional by Dunnet (1976), who delineated a series of 'spoon' faults in the area immediately north of Mt Isa, and as thrusts by Bell (1983), who applied a duplex model to the same area. Similar structures in an area 60 km south-southeast of Mt Isa are interpreted by Passchier (1986) as brittle extensional detachments. Dahlstrom (1970) and Coward (1982) have emphasized the similarities between low angle thrust zones and low angle normal faults: both tend to have staircase trajectories (flats and ramps) and both may develop as imbricate sequences. Passchier (1986) pointed out the differences, concluding that stratigraphical repetitions and persistent parallelism over large areas of bedding planes and faults may distinguish thrust terranes from extensional terranes.

These two prerequisites of thrusting are satisfied in the eastern part of the Mary Kathleen 1:100 000 Sheet area, which lies roughly equidistant between Bell's thrust terrane and Passchier's extensional terrane. Based on stratigraphical repetitions and the presence of bedding-parallel phyllonites, thrusts can be inferred at a number of stratigraphic levels. The thrusts are marked by zones of brecciation and/or talc-bearing phyllonites, both of which are transected by \( S_2 \).

Thrust sheets are commonly characterized by variation of strain which may result in the transition of folds with axial planes at right angles to the thrust plane into isoclinal folds with axial planes almost parallel to the thrust plane (Sanderson 1982). In this area, a single thrust sheet possesses a continuous series of sigmoidal pre-\( D_2 \) foliations similar to those calculated by Sanderson (1982, fig. 6b) for thrust sheets with layer-parallel shortening and similar to those documented by Mitra and Elliott (1980) from a thrust system in the southern Appalachians. Spaced axial-plane cleavages of upright folds within the thrust sheets are, when followed downwards, asymptotic to the thrust plane and become phyllonitic in the immediate vicinity of the thrust plane (Fig. 4). The phyllonic layering (\( S_{1A} \), Fig. 5) ranges from a very finely spaced, dominal, differentiated layering to a slaty cleavage; the \( D_1 \) extension lineation, \( L_1 \) is well developed on the layering and is defined by prolate quartz aggregates and aligned micaceous minerals. \( L_1 \) trends east–west. West-verging folds accompany parts of the thrust as in the uppermost Corella Formation west of the Deighton Syncline (e.g. GR 830057; point F in Fig. 1) or a low angle stylolitic or fracture cleavage may be confined to the vicinity of the nappe contact, increasing in intensity towards it, as in the Magna Lynn Metabasalts near the Magna Lynn Mine. Axial-plane foliations of
Fig. 4 Schematic representation of the sigmoidal foliation development in a D1 thrust sheet. S1c, in the low-strain domains within the thrust sheets, rotates via S1b, a shear foliation close to the thrust planes, towards the phyllonitic S1a, which is (sub-) parallel to bedding at the thrust plane itself. Folds become progressively more non-cylindrical towards the thrust plane. Lateral complexities (e.g. tear faults) have been omitted.

these D1 structures, named S1b, can be explained in terms of differential movement along both sides of the thrust (Figs 4, 6). This zone of D1b folds can be very narrow as for example southeast of the Deighton Syncline; in other places it is obscured. When traced upwards into the centre of the thrust sheets, the cleavage becomes an axial-plane foliation (locally a spaced cleavage) associated with open upright northeast-trending folds. This foliation is called S1c (Fig. 4). The folds probably reflect shortening approximately parallel to the thrusting direction. Thrusting, folding and the development of the continuous and sigmoidal series of S1 orientations are interpreted as coeval.

The thrust geometry of the northeastern part of the Mary Kathleen Sheet, northwest of the abandoned Mary Kathleen township, can be demonstrated from two key areas. The first, the 'West Leichhardt Klippe', is interpreted as a folded imbricate stack of thrust sheets above a floor thrust, and the second, the 'Deighton Klippe', as a doubly plunging synclinal klippe above a roof thrust (Fig. 1). The partial duplexes of the West Leichhardt Klippe and Mt Devine/Deighton area do not necessarily belong to the same duplex: a local connotation only need be attributed to the terms roof thrust and floor thrust.

WEST LEICHHARDT KLIPPE

This structure is located north of 'West Leichhardt' station in the adjoining parts of the Mary Kathleen and Prospector 1:100 000 Sheets (Figs 1, 7). Because numerous younger faults obscure the klippe in the Prospector Sheet area, our description is based on the southern part of the klippe in the Mary Kathleen Sheet, where a north-south elongate body of metasediments is surrounded by the Kalkadoon Granite. The sediments consist of a fining-upwards sequence of conglomerate, quartzite (sandstone) and siltstone mapped by Derrick et al. (1977) as the Surprise Creek Beds. This sequence is repeated at least seven times, twice in its entirety, the remainder involving only parts of the succession. Quartz-filled faults and breccias mark the contacts between the repetitions. Persistent parallelism of the faults and bedding planes suggests thrusting.

At the contact between the Surprise Creek Beds and the Kalkadoon Granite there is a phyllonitic foliation parallel to bedding. Within 50 m of the sediment/granite boundary, the contacts between the stratigraphic repetitions rotate towards parallelism with the phyllonitic layering. This curvature was studied in detail at the mid-eastern boundary of the West Leichhardt Klippe, where a continuous series of intensely deformed conglomerate, sometimes less than 10 m thick, is juxtaposed against the boundary phyllonites. In one place (GR 662300; point G in Fig. 1) isoclinal, southeast-plunging, inclined folds with attenuated lower limbs are present in sediments immediately above the phyllonitic layering. The layering probably originated from deformation along a D1 (floor) thrust. The southeasterly plunging isoclinal folds, associated with this floor thrust, are D1b folds. Above the floor thrust, in the meta-sediments, thrusting was more brittle and quartz-filled faults and breccias developed between a north-dipping pile of imbricates.

The mass of the Surprise Creek Beds appears to represent a folded klippe. Subsequent to the D1 thrusting, east-west shortening of the thrust complex produced in a large, north-plunging, upright syncline (schematically represented in Fig. 8). An axial-plane foliation (S2) cross-cutting the imbricates is penetratively developed as a spaced cleavage in the quartzites and as a slaty cleavage in the siltstones. Locally north-south upright D2 shear zones with vertical extension lineations reactivated the floor thrust (e.g. GR 663282; point H in Fig. 1). The klippe was later openly refolded around a vertical east-west-trending D3 fold.
Fig. 5  
$D_2$ crenulations on a $D_1$ phyllonite, $S_{1A}$ (GR 988863; point E in Fig. 1). The matchstick is parallel to $L_1$.

Fig. 6  
$D_2$ high-strain zone in incompetent beds of the upper Corella Formation. Axial planes of the highly non-cylindrical folds are almost parallel to bedding. In the low-strain zone adjacent to it, axial planes are vertical and parallel to the regional $S_2$ (GR 830057; point F in Fig. 1).
The Deighton Syncline, located north of the Barkly Highway in the eastern Mary Kathleen 1:100 000 Sheet area (Figs 1, 9), falls into the same tectonostratigraphic zone as the Alligator Syncline, interpreted by Passchier (1986) as an extensional zone. The Deighton Syncline is a doubly plunging north-south elongated syncline. Its core is occupied by the Mt Albert Group (Deighton Quartzite and White Blow Formation), whereas it is surrounded by the Corella Formation of the Mary Kathleen Group.

Northwest and west of the Deighton Syncline, in the Mt Devine fault system (Fig. 1), are steep, north-trending, eastward-younging repetitions of...
Fig. 9 Structural map of the Deighton Klippe and surroundings.
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Fig. 10 Model of an extensional imbricate stack, developed in the Mary Kathleen Group. In this model (which is not that favoured in this paper) the Deighton Quartzite (Ppd) represents a rift phase and is deposited in small, rapidly subsiding sub-basins, while the White Blow Formation (Ppw) represents the subsequent sag phase. However, two prerequisites of this model, the non-parallelism between the normal (growth) faults and bedding and the non-tectonic contact between the Mary Kathleen Group and the Mt Albert Group, prove to be unsatisfiable. Abbreviations as in Fig. 1.

Fig. 11 Model of constrictional imbricate stack, developed in the Mary Kathleen Group (MKG). (a) The Mt Albert Group (MAG) could either be folded around blind thrusts (I) in the underlying Mary Kathleen Group or could overlie a roof thrust ABC. (b) This geometry, which is in better accordance with the observed field relations, can be explained either by erosion of the imbricate stack and subsequent deposition of the Mt Albert Group, or by an 'overstep' mode of thrust propagation in which a younger thrust at the base of the Mt Albert Group truncates the imbricate stack. Only in the latter case will AC be a tectonic contact. Abbreviations as in Fig. 1.

metavolcanics of the Argylla Formation, the Ballara Quartzite and calcisilicate rocks of the Corella Formation. The contacts between these lithological repetitions are marked by breccias and bedding-parallel phyllonites with east-plunging stretching lineations. Without further information, such a geometry might be explained either by a folded stack of listric normal faults, originally dipping towards the west (Fig. 10), or by a folded stack of thrust imbricates, originally dipping towards the east (Fig. 11). In the former interpretation, the Mt Albert Group could have been deposited in small rapidly subsiding asymmetrical basins on top of the listric faults and would unconformably overlie the Mary Kathleen Group (Fig. 10). Such an interpretation would fit tentative models proposed by Etheridge et al (1985) and Blake et al (1985). These workers envisaged the evolution of the inlier as involving three early intracratonic extensional phases during which rift (and sag) sediments were deposited. The Mary Kathleen Group would represent the second sag phase. Following their line of reasoning, the Mt Albert Group may represent the third.

To distinguish extensional from shortening (i.e. thrusting) mechanisms for the formation of sequential repetitions, it is necessary to examine: (1) the nature of the contact between the two groups and (2) the relation of the listric faults to bedding within the Mt Devine Fault System. In the latter, although D2 considerably steepened the faults, phyllonites developed on the faults dip consistently towards the east and are (sub-)parallel to bedding over large areas; in addition there is no thinning or excision of rock units, a feature generally produced by extension. Hence, we conclude that the Mt Devine faults constitute a thrust system. However, the mass of Mt Albert Group material on top of the Mt Devine system is not transected by the thrusts. The question therefore arises as to whether the Mt Albert Group is autochthonous and was deposited after the thrusting, or whether this particular part of the Mt Albert Group is allochthonous, either having been transported passively over a relatively young roof thrust or having decapitated and overstepped the Mt Devine Imbricate Stack during thrust propagation (Fig. 11).

An elucidating traverse in this respect can be followed from GR 830057 to GR 830069 (respectively points F and I in Fig. 1), through the upper member of the Corella Formation, ending at the
base of the Deighton Quartzite on the western limb of the D2 syncline (Fig. 12a, b). The contact between these two formations has been interpreted as an unconformity (Derrick et al. 1977). At the base of the traverse west-verging folds and small-scale thrusts are developed in bands of alternating incompetent calcareous and competent calcisilicate material. The incompetent calcareous layers are affected by a layer-parallel shear strain. Fold axes are rotated towards an east-plunging orientation, giving rise to highly non-cylindrical folds with both south- and north-verging fold extremes. The folds commonly are nearly isoclinal, and attenuated lower limbs are the rule rather than the exception. Axial planes in these beds are almost always layer-parallel, that is dipping 40-50° towards the east (Fig. 6). In strong contrast, axial planes in the competent beds are vertical north-trending. These upright folds can be traced upwards into crenulations within 20 m of mica schist, present just below the platy base of the Deighton Quartzite (platy for 0.1-1 m). Since these crenulations overprint recumbent folds with sheared out lower limbs, and because they can be correlated with the regional D2 fold pattern, we suggest they are D2 folds. Given that the crenulations are D2 folds, then the folds...
described previously within the upper Corella Formation are also $D_2$ structures. The unusual geometry wherein $S_2$ in the competent layers, rather than $S_2$ in the incompetent layers, is parallel to the axial plane of the large structure, can be explained by layer-parallel (flexural) flow in the incompetent beds during folding, analogous to folds described by Bayly (1965).

The mica schists and the related sheared out recumbent folds (west-verging) are related to differential movement with the upper block moving towards the west. The west-verging $D_1$ folds are not restricted to the western limb of the Deighton Syncline, but are also present on the southeast limb of the 'Campbell Klippe' (Fig. 1), for example, at GR 845961 (point K in Fig. 1). This excludes a $D_2$ flexural slip mode of development for the west-verging folds. This interpretation of the kinematics is supported by a progressively stronger development of an east-west extension lineation, defined by prolate quartz aggregates and aligned micaceous minerals, towards the top of the mica schist zone. The contact between the Mt Albert Group and the Mary Kathleen Group, then, on top of the mica schists, is marked by a phyllonitic layering, parallel with and overlain by platy Deighton Quartzite with a weak east-plunging extension lineation. This strongly suggests that the Mary Kathleen Group and the Mt Albert Group are separated by a tectonic contact.

In summary, traversing from the upper part of the Corella Formation into the Deighton Quartzite, there are firstly only $D_2$ structures, then a thin mica schist zone, with an upwards progressively stronger development of east-west extension lineations and west-verging folds, which are crenulated by $D_2$ folds; above this lies a crenulated phyllonite, again crenulated by $D_2$, and overlain by the platy basal unit of the Deighton Quartzite (Fig. 12a). The contact between the Mt Albert and Mary Kathleen Groups is therefore tectonic and the possibility that the Mt Albert Group was deposited in place after the thrusting event is discounted. The schistosity is interpreted as $S_{1A}$, the west-verging folds as $D_{1B}$ folds, the mica schist zone as marking a major thrust zone, and the extension lineation as $L_1$.

The repetition of the Argylla Formation, Ballara Quartzite and Corella Formation west and northwest of the Deighton Syncline can now be interpreted as an imbricate stack of $D_1$ thrust nappes (Fig. 1). Since the imbricates are truncated and overlain by the major thrustplane at the base of the Mt Albert Group they are interpreted as footwall imbricates. The mass of Mt Albert Group material in the core of the Deighton Syncline overlies a roof thrust and should be regarded as a tectonic rather than a stratigraphic entity; it will therefore be called the Deighton Klippe. Thus, the picture of a partial duplex emerges (terminology after Dahlstrom 1970; Butler 1982; Boyer & Elliott 1982). Gypsum within the black shales of the uppermost Corella Formation recovered from BMR Cloncurry No. 5 drillhole (Hill & Duff 1975), may have facilitated thrusting. The presence of marialitic scapolite (e.g. Derrick et al 1977) and the geochemistry and field relations of the cordierite-anthophyllite rocks (Reinhardt 1986) within the Corella Formation point towards an evaporitic origin. The low shear strength of the well-bedded evaporites would have made the Corella Formation an unrivalled candidate for the development of $D_1$ thrust-planes.

Derrick et al (1977) interpreted the contact between the Mt Albert Group and the Mary Kathleen Group as a stratigraphic unconformity. Considering, however, the gross lithological resemblance of the two groups and the inferred thrust plane between them, we propose the alternative interpretation that the Mt Albert Group is a stratigraphic equivalent of the Mary Kathleen Group. East-west extension lineations and west-verging $D_{1B}$ folds indicate an eastwards origin for the now allochthonous Mt Albert Thrust Nappe.

The Blockade Block (Fig. 1), consisting mainly of Leichhardt Volcanics, is located northeast of the Deighton Klippe and separated from it by a narrow strip of upper Corella Formation. Prominent northwest-trending shear zones, considered to postdate $D_2$ folding (Derrick et al 1977), transect this block, but from their orientation and analogy with the Mt Devine imbricate thrusts, some could be explained as $D_1$ thrusts at a deeper level.

Approximately northeast-trending upright folds are present in the Mt Albert nappe and in the Mary Kathleen Group; the former is characterized by a spaced or slaty axial-plane cleavage. In our view these folds result from shortening (sub-parallel to the local $D_1$ thrusting direction) and should be classified as $D_{1C}$ folds (Fig. 4).
Younger, north-trending, upright folds overprint all early structures, locally crenulate the phyllonitic layering (Fig. 5) and cause fold overprinting patterns. These are best displayed in the White Blow Formation, in the centre of the Deighton Klippe (e.g. GR 825104; point L in Fig. 1) and around GR 852097 (point M in Fig. 1) in the upper Corella Formation. This later folding, often accompanied by a vertical north—south axial-plane fracture cleavage, is highly non-cylindrical and is typical of D$_2$ folding over the entire Mt Isa Inlier. This non-cylindrical geometry results in the map-scale alignment of doubly plunging folds. The present erosion level reveals elongated and aligned klippes of Mt Albert nappe material (Fig. 12c).

**D$_1$-NAPPE MOVEMENT DIRECTION**

The imbricates of the West Leichhardt Klippe are symmetrically folded around a north-trending D$_2$ axial plane. Because the fold axes consistently plunge towards the north, the imbricates are likely to have been north-dipping before D$_2$, and because thrusts tend to climb in the direction of movement, this indicates a southward movement direction. A northern lateral ramp geometry on a thrust system with an east—west component in the movement direction is not an alternative because it does not explain the intersection of all imbricate thrusts with the floor thrust at both limbs of the D$_2$ syncline.

In contrast to the north-dipping imbricates of the West Leichhardt Klippe, the thrust imbricates of the Mt Devine system are all steeply east-dipping. This could be explained in two ways: either by westward moving and climbing imbricates, that have been further steepened by D$_2$, or by an easterly lateral ramp of a north—south moving nappe system, also steepened by D$_2$. In the latter case, displacements would have been large: the overlap of Ballara Quartzite in the north—south direction in the Mt Devine Imbricate Stack area amounts to more than 50 km. In the former case, considerably less displacement is needed to achieve the observed geometry. No reliable sense of shear indicators have been found in the Mt Devine Imbricate Stack.

More information can be obtained from the roof thrust at the base of the Deighton Quartzite, where L$_1$ is consistently down-dip on the phyllonitic layering, S$_{1A}$, respectively, east-plunging on the western limb of the Deighton Klippe and west-plunging on the eastern limb. West-verging shear folds (D$_{1B}$) are developed in the upper Corella Formation under the roof thrust. Since east—west pre-D$_2$ lineations and west-verging shear folds have also been found in the same tectonostratigraphic setting in large D$_2$ pressure shadows at the southern margin of the Campbell Klippe (just south of the Deighton Klippe; Fig. 1), a D$_2$ reorientation of these structures due to flexural slip or flow can be excluded. This means that the lineations reliably indicate direction of shear and that the west-verging folds indicate sense of shear, in this case pointing to a westward translation for the Deighton Klippe. Southwest-trending D$_1C$ folds in incompetent rock units of the upper Corella and White Blow Formations suggest southeast—northwest shortening during D$_1$. Thus, we conclude, that thrust movements in this region are most likely to have been towards the west or northwest.

**DISCUSSION**

Our interpretation opposes the idea that the Mt Albert Group was deposited in small basins on top of a horizontally extending Mary Kathleen Group during a period of intracratonic extension, but argues in favour of a pre-D$_2$ thrusting event. This hypothesis does not exclude a phase of extension predating the thrusting event, and lithospheric extension can be used to explain the subsidence responsible for accumulation of the various cover sequences. A thinned crust may also have led to a concentration of deformation in these zones.

According to gravity and refraction seismic data (Dooley 1980; Finlayson 1982), the present crustal thickness at the Mt Isa Inlier is probably 40–45 km. Post-metamorphic erosion of large areas of the inlier to the currently exposed greenschist facies level has removed 10–15 km, so that the crustal thickness immediately after D$_2$ amounted to $55 \pm 5$ km. An estimation by the authors of 50% east—west shortening for D$_2$ (under plane strain conditions) would imply a doubling of the thickness of the crust, leaving a pre-D$_2$ crustal thickness of 28 km.

Syn-D$_2$ andalusite and cordierite-anthophyllite assemblages at the base of the Corella
Appendix 1

Formation (Reinhardt 1986; Passchier 1986) with rare staurolites (according to Derrick et al. 1977, staurolite assemblages are confined to the White Blow Formation) point towards bulk geothermal gradients of approximately 35°C/km at a pressure of approximately 3.5 kbar (Hietanen 1959, 1967; Holdaway 1971), implying a depth of approximately 14 km (assuming a bulk density of the cover of 2.5 g/cm³). Even if thrusting during D₁ locally thickened the cover by 100% (without thickening the lower crust), this would still mean that a stratigraphic pile 7 km thick was deposited above the base of the Corella Formation (assuming a 40 km thick pre-extension crust, this results in an extension factor of 2). The Corella Formation could, like other sag-basin deposits, possibly have had this thickness, but no such successions exist in the central Mt Isa Inlier.

A second possibility is that a later extensional phase in the Mt Isa Inlier (Blake et al. 1985) resulted in the deposition of the Mt Isa Group (Blake 1987), a sag-phase unit presently exposed only in the western part of the inlier and with an inferred age of 1670 Ma. A recent numerical modelling study by Voorhoeve and Houseman (1986) shows that the Mt Isa Group is unlikely to have been deposited during the thermal re-adjustment of the rifting phase during which the Argylla Formation (1777 ± 7 Ma, Page 1978) was deposited, because sagging of the lithosphere is a decay process with a half-life of only 60 Ma. In the central Mary Kathleen area there are no young rift sediments overlying the Mary Kathleen Group. Thus, either the young rift with its faulted margin(s) is completely eroded in this area, or the rift was developed west of the central Mary Kathleen Sheet area, and is represented by remnants of the Fiery Creek Volcanics and Carters Bore Rhyolite (1678 ± 1 Ma, Page 1978). The latter interpretation would explain the absence, or rarity, of extensional faults affecting the Mary Kathleen Group in the central Mary Kathleen area. During laterally extensive subsidence, associated with the rifting event at 1678 Ma, the Mt Isa Group may have been deposited on top of the Mary Kathleen Group in this area.

CONCLUSIONS

Thrusts in the area are marked by zones of brecciation and/or phyllonites, both of which are transected by D₂ structures; the thrusts are therefore developed during the early deformation stage (D₁).

After emplacement on a roof thrust, overlying an imbricate stack, the Mt Albert nappe (D₁) was folded into elongate domes and basins (D₂). Subsequent erosion produced aligned, elongate klippen of the Mt Albert nappe, between which the Mt Devine and Blockade imbricate slices are exposed. The floor thrust beneath the imbricates does not crop out. To the west, the West Leichhardt Klippe comprises a folded floor thrust together with overlying imbricate slices, but there is no reason to assume that this is the same floor thrust supposedly underlying the Deighton roof thrust. This means that no complete duplex can be observed, and that only isolated hanging-wall imbricates (plus floor thrust) and footwall imbricates (plus roof thrust) can be reconstructed with certainty.

D₁ movement directions are not the same on all thrusts: a westwards or north-westwards direction of displacement is indicated for the Deighton roof thrust, whereas a strong southerly directed movement vector can be deduced for the West Leichhardt Klippe.

Thermal relaxation after rifting at 1678 Ma in the western part of the inlier may have resulted in the accumulation of the Mt Isa Group sediments in a sag-basin, but there is no evidence that this young extensional activity affected the central Mary Kathleen area.

Given the lithological similarity to the Mary Kathleen Group rocks and the structural evidence for thrusting it is reasonable to conclude that the Mt Albert Group is a lateral sedimentary facies equivalent to the Mary Kathleen Group rather than a younger stratigraphic repetition of similar lithology.

ACKNOWLEDGMENTS

This project was funded by the Dutch Molengraaff Funds, the Dutch Schürmann Funds, CRA Exploration Pty Ltd and the University Funds of the State University, Utrecht, The Netherlands. Logistical support in the field was provided by the Bureau of Mineral Resources, Geology and Geophysics. During the 1985 field season RJHL held a PhD scholarship from the Australian National University. To all these organizations, we express our gratitude. Toni Convine (BMR) did the drafting. We are particularly thankful to
Mike Etheridge, who brought the possibility of thrusting in the Deighton area to our attention, and to Gordon Lister, who supervised this project. We are also indebted to Mike Rickard and Rod Holcombe for their positive reviewing of earlier drafts. This work has been carried out under the auspices of the BMR Mount Isa Regional and Tectonic History Project.

REFERENCES


SANDERSON D. J. 1982. Models of strain variation in


(Received 14 April 1986; accepted 24 October 1986)
Extension tectonics extended: new results from the Mount Isa Inlier

Further results from the Mount Isa Inlier of northwest Queensland considerably enlarge the area of known extensional tectonics, and reinforce the importance of regional extension at a significant stage in the geological stratigraphy of the Inlier: the extensional structures may have initiated the basins in which the Mount Isa type strataform zinc-lead-silver orebodies were deposited, and could also have exerted an important control on the mineralisation.

Extensional origin of the Deighton Klippe

Since the current BMR work in the Mount Isa Inlier began in 1983, the importance of major low-angle extensional faulting accompanied by large-scale subhorizontal movement of rock masses has become increasingly recognised. Dunnet (1976: Philosophical Transactions of the Royal Society of London, Series A, 283, 333–344) first suggested the possibility of low-angle normal faulting when he proposed that the Hilton and Mount Isa orebodies were originally parts of a single mass, later separated by major low-angle extensional faulting. Pashcier (1986; Geology, 14, 1008–1011) and BMR (1987; Research Newsletter 6, 10–11) respectively documented early low-angle normal faulting in the southern part of the Alligator Syncline and in the Bulonga Anticline (Fig. 2). Loosveld & Schreurs showed that the Deighton Klippe is in tectonic contact with the underlying Corella Formation, weaken away from the contact, and in many places are deformed by a steep, northerly-striking foliation (S2).

Bedding-parallel normal faults filled with metadolerite about 20 m thick at the base of the Deighton Quartzite, thereby bringing the overlying sandstone into contact with the Corella Formation.

Mete-scale normal fault blocks cut by S2 in the Deighton Quartzite; similar centimetre-scale normal fault blocks are cut by S2 in the Scorpion outlier.

Younger rocks placed tectonically over older

The evidence indicates west-directed extensional faulting, brecciation, and tectonisation along a system of pre-D2 west-dipping low-angle normal faults (Fig. 4a). The bedding-parallel fault beneath the Deighton Klippe was originally a flat that formed part of this system. The normal faults were probably pre-D1 in origin, because high and low-angle normal faults filled with metadolerite containing S1 foliation, stretching lineation, and metamorphic mineral assemblages occur in the Bulonga Anticline (BMR, 1987), the Little Beauty Syncline (P.R. Williams, unpublished results), and in the Wonga Belt (P.J. Pearson, personal communication). The extension (Fig. 4b) placed younger rocks over older, i.e., Deighton over Corella and Corella over Leichhardt. Extension was followed by north-south compression (D1), which formed the F1 synform in the White Blow Formation of the Deighton Klippe, and then by east–west D2 compression (Fig. 5). This produced minor thrust-faulting and folding on the western side of the Deighton Klippe, and then the major north–south trending upright folds (such as the synform forming the Deighton Klippe) and steep north–south striking axial-plane foliation (S2).

Of the other Deighton Quartzite outliers, the Charley Creek outlier has a fault right around it, and is identical to the Deighton Klippe. In contrast, the Campbell, Scorpion, and Maylene outliers are in conformable sequence with the underlying Corella Formation on their western sides (including the passage bed interval described above), and are in steep fault contact with older rocks (Leichhardt, Magna Lynn, or Ballara) on their eastern sides. At the Scorpion outlier, the fault is marked by prominent breccia which is cut by and therefore pre-dates S2 foliation. The eastern bounding faults at the Campbell and Maylene outliers are syn or post-S2. Maylene has a steeply east-dipping thrust-fault coupled with sinistral strike-slip, and Campbell has a vertical to steeply west-dipping normal fault. The structural setting of the Scorpion and Maylene outliers is shown in the appropriate parts of Fig. 5. The folding rotated the early normal faults, thereby reversing their dip and sense of throw and making them appear to be thrust faults. This removes the necessity for west–east-directed thrusting in the Deighton area, as suggested by BMR (1985: Research Newsletter 2, 11) and by Loosveld & Schreurs (1987). On the eastern side of the Campbell outlier, undeformed fault breccias cut S2, separate different orientations of S2 on each side of the fault, and in places contain flattened clasts of S1-foliated rock in random orientations. This indicates post-D2 reactivation of the early fault, and similarly for the combined thrust and strike-slip fault on the eastern side of the Maylene outlier.

The phenomena described here reinforce the importance of the regional extension that postdated deposition of the Mary Kathleen and Mount Albert Groups. We suggest that it was this crustal extension that initiated the basins in which the Mount Isa Group and its related and correlated sequences were deposited. The extensional structures may have localised the Mount Isa-type zinc-lead-silver mineralisation, similarly to the syn-sedimentary normal faults which allowed metal-bearing fluid to enter the McArthur Basin to the north, to form the HYC zinc-lead-silver deposit.
Appendix 1

Fig. 2. Generalised geological map of central Mount Isa Inlier. AS Alligator Syncline, BA Belonga Anticline, BB Blockade Block, CO Campbell Outlier, CCO Charleyn Creek Outlier, DK Deighton Klippe, LBS Little Beauty Syncline, MO Maylene Outlier, SO Scorpion Outlier, UO Unnamed outlier, WB Wonga Belt, WLK West Leichhardt Klippe.

Fig. 3. Geological map of Deighton Quartzite outliers. A–B indicates position of section line in Fig. 5.

Fig. 4. System of hypothetical normal faults together with a bedding-parallel fault (a) before, and (b) after extension.

Fig. 5. Schematic cross-section through Fig. 3, showing normal faults fold by D2, Deighton Klippe, and Scorpion and Maylene outliers projected on to line of section.
"Reply" to "Discussion" by Ramon J.H. Loosveld

The discussion by Stewart raises a number of principal problems, related to thrust tectonics, which affect a previously thinned upper crust. In our paper, we have compared the pre-D$_{c2}$ geometry of the Deighton Syncline and its surroundings with both contractional and extensional geometries, and concluded that the Deighton Syncline is a D$_{c1}$ thrust klippe folded around a D$_{c2}$ synform. We put forward the theory that in the Deighton/Mount Devine area, the Ballara Quartzite and Corella Formation are separated by a roof thrust from the lithologically similar Deighton Quartzite and White Blow Formation, and could very well be lateral correlatives of these. Stewart, although acknowledging W-directed D$_{c1}$ thrusting at the base of the western side of the Deighton Klippe, disputes the possibility of such a correlation and suggests that the imbrication within the Mount Devine area (immediately west of the Deighton Syncline in Stewart's figure 5) is caused by (the rotation of) early normal faults.

A diagnostic difference between the two opposite fault regimes is that in the case of normal faults, the dip direction of the fault is originally opposite to the dip direction of bedding, whereas in the case of thrusting, the faults and bedding planes dip towards the same direction. In the Mount Devine Imbricate Stack, between the Deighton Klippe and the Scorpion Outlier (Stewart's figure 5), both the bedding planes and the faults dip towards the east. Stewart's observation is almost the same: the "ground level" in his figure 5 shows east-dipping and vertical faults. There is no evidence on the surface to suggest that the faults are curved and change orientation through the vertical (the XY-plane of D$_{c2}$) and dip westwards both underground and in the now eroded higher topographic levels, as in Stewart's figure 5. The faults of the Mount Devine Imbricate Stack, which are generally (sub-)parallel to bedding, therefore have to be interpreted as thrusts. These thrust planes pre-date D$_{c2}$. Complementing mesoscopic sense of shear indicators, the N-S trend of the imbricate stack also points to at least a component of W-directed thrusting.

Much of Stewart's discussion is based on observations on the contact between the Corella Formation and the Deighton Quartzite. Stewart argues for the presence of "passage-beds" between these two units. However, it can be extremely difficult in the field to distinguish a truly transitional sequence from a series of displaced and isolated tectonic units, as e.g. rootless folds, boudinaged segments or thrust slices. It is even more difficult, when, as in our case, most D$_{c1}$ fabrics are recrystallized during a later pervasive deformational event. The lower limbs of the recumbent folds at the base of the Deighton Quartzite e.g., which have been observed by Stewart, Bettess (1987) and myself (unpubl. data), are boudinaged and brecciated. Such structural features may well
reflect local orientation-dependent extension due to large shear strains at the contact between the two lithologies, and are not necessarily an indication of regional extension, as suggested by Stewart.

The same argument applies to Stewart's other observations of the Deighton Quartzite/Corella Formation contact. We interpret a roof thrust separating the Deighton Quartzite in the Deighton Klippe from the underlying Corella Formation in the highest exposed thrust sheet in the Mount Devine Imbricate Stack. Stewart, on the other hand, suggests that an extensional fault, reactivated by minor thrust displacements, separates the two. Stewart has rightly pointed out (written comm., 1987) that the W-verging folds on the contact (described by Loosveld & Schreurs) are not diagnostic for either regime, but are only indicative of a movement vector. Inconsistently, he then does attribute brecciation and small-scale normal faulting to regional normal faulting. Both of these, however, are, like the boudinaged quartzites and the W-verging faults, only indicative for a zone of high shear strain, independent of the sense of displacement along the shear zone.

Stewart postulates a major extensional phase overprinted by a weak (pre-)Dc2 thrusting. All of his observations, though, can be explained by thrusting alone, except the omission of the Magna Lynn Metabasalt, the Argylla Formation and the Ballara Quartzite, northeast of the four Deighton Quartzite outliers. Regionally, however, the number of litho-stratigraphic repetitions outweighs these four excisions. It is not justified to base an all-embracing extensional model on local structures, indicating local extension. It also remains to be proven that these excisions predate Dc2, as in some cases Dc3 strike-slip faulting may play an important role.

The Loosveld & Schreurs and Stewart viewpoints are not diametrically opposed. The geometry of the Deighton area, let alone the kinematics and dynamics, is not yet fully understood, and early extensional structures, local and regional, are not excluded by the thrust model. However, although extensional tectonics have played an important role in the early (pre-1680 Ma.) evolution of the inlier, the bulk of the early structures here (those generated between 1680 and 1550 Ma.) fit a contractional rather than an extensional regime. Despite Stewart's new data and comparisons with other outliers of Deighton Quartzite, we still believe, that, as stated in the abstract of the original paper, "the possibility that a major tectonic repetition rather than sedimentary repetition occurs in the cover sequences cannot be excluded at present". This tectonic stacking of litho-stratigraphical units predates Dc2 and postdates Dc1 (as it affects the Argylla Formation, the Ballara Quartzite and the Corella Formation), and, by tentative correlation with thrusting north of Mount Isa township in Mount Isa Group lithologies (Bell, 1983), probably also postdates Dc3 (1680-1670 Ma.). Page & Bell (1986) dated Dc1 in the
Sybella Batholith at 1610±13 Ma., which fits these structural constraints. Similarly bracketed contractional pre-D$_{c2}$ structures have been recognized in all four tectono-stratigraphic domains studied in this Ph.D. project (the Deighton and West-Leichhardt Klippen, the Mitakoodi Anticlinorium, the Marimo-Staveley Block, and the central Soldiers Cap Group belt). This pervasive D$_{c1}$ event could well have triggered convective "erosion" (=thinning) of the thickened mantle lithosphere, causing prograde low-P facies conditions at D$_{c2}$ times.
APPENDIX 2

ABBREVIATIONS, NOTATIONS AND ASSIGNED VALUES

LOOSVELD

In condensed form submitted with Chapter 4 for publication in "Tectonophysics"
### Abbreviations, Notations and Assigned Values

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ac</td>
<td>accuracy</td>
</tr>
<tr>
<td>Agr</td>
<td>radiogenic heat production of granite</td>
</tr>
<tr>
<td>Ao</td>
<td>radiogenic heat production at surface</td>
</tr>
<tr>
<td>BDT</td>
<td>brittle/ductile transition (only in code TR2Bf)</td>
</tr>
<tr>
<td>C</td>
<td>concentration</td>
</tr>
<tr>
<td>Cf</td>
<td>specific heat capacity of fluid</td>
</tr>
<tr>
<td>cm</td>
<td>specific heat capacity of solidifying magma, corrected for latent heat of fusion</td>
</tr>
<tr>
<td>cs</td>
<td>specific heat capacity of solid</td>
</tr>
<tr>
<td>CRU</td>
<td>thickness of undeformed crust</td>
</tr>
<tr>
<td>diff</td>
<td>thermal diffusivity of solid</td>
</tr>
<tr>
<td>D</td>
<td>radiogenic-heat scale length (A = Ao exp(-D))</td>
</tr>
<tr>
<td>E</td>
<td>upwards velocity (here, = erosion rate)</td>
</tr>
<tr>
<td>Ed</td>
<td>thickness of finally eroded package</td>
</tr>
<tr>
<td>f</td>
<td>the progressive amount of thickening; f = ( \dot{e} t ), in case of constant strain rate</td>
</tr>
<tr>
<td>f_end</td>
<td>the finite amount of thickening</td>
</tr>
<tr>
<td>h</td>
<td>1/( \text{imax} )</td>
</tr>
<tr>
<td>HFU</td>
<td>heat flow unit; ( 1 \text{HFU} = 10^{-6} \text{cal/cm}^2\text{s} = 41.84 \text{mW/m}^2 )</td>
</tr>
<tr>
<td>HPE</td>
<td>heat producing elements (mainly U, Th, K)</td>
</tr>
<tr>
<td>imax</td>
<td>total amount of depth steps</td>
</tr>
<tr>
<td>jmax</td>
<td>total amount of timesteps (j=1/k)</td>
</tr>
<tr>
<td>k</td>
<td>1/( \text{imax} ) (k=( rh^2 ))</td>
</tr>
<tr>
<td>K</td>
<td>thermal conductivity of solid</td>
</tr>
<tr>
<td>L</td>
<td>lithospheric thickness</td>
</tr>
<tr>
<td>L_r</td>
<td>latent heat of fusion of magma</td>
</tr>
<tr>
<td>mtdet</td>
<td>time in number of timesteps (( \Delta t )) until crust-mantle detachment</td>
</tr>
<tr>
<td>mtend</td>
<td>time in number of timesteps (( \Delta t )) until end of thickening</td>
</tr>
<tr>
<td>mthm</td>
<td>time in number of timesteps (( \Delta t )) until end of &quot;hot mode&quot; (convection)</td>
</tr>
<tr>
<td>M</td>
<td>( \frac{(\rho_c c_p)}{(\rho_s c_s)} = 1 )</td>
</tr>
<tr>
<td>MKFB</td>
<td>Mary Kathleen Fold Belt</td>
</tr>
<tr>
<td>r</td>
<td>stability factor; ( r = k \Delta t / \Delta x^2 = k / h^2 )</td>
</tr>
<tr>
<td>SCG</td>
<td>Soldiers Cap Group</td>
</tr>
<tr>
<td>SR</td>
<td>absolute exponent of strain rate (e.g. if ( \dot{e} = 10^{-14} ) then SR=14)</td>
</tr>
<tr>
<td>t_det</td>
<td>time of crust-mantle detachment during progressive thickening</td>
</tr>
<tr>
<td>Tempdet</td>
<td>temperature at the base of the thickened crust</td>
</tr>
<tr>
<td>t</td>
<td>time in seconds</td>
</tr>
<tr>
<td>Variable</td>
<td>Description</td>
</tr>
<tr>
<td>----------</td>
<td>-----------------------------------------------------------------------------</td>
</tr>
<tr>
<td>$t_{end}$ ($=t_{fend}$)</td>
<td>time at end of thickening</td>
</tr>
<tr>
<td>$t_{hrm}$</td>
<td>duration of &quot;hot mode&quot; (convection) at the base of the thickened crust</td>
</tr>
<tr>
<td>$T_i$</td>
<td>intrusion temperature</td>
</tr>
<tr>
<td>$T_l$</td>
<td>liquidus temperature</td>
</tr>
<tr>
<td>$T_{max}$</td>
<td>temperature at the base of the lithosphere</td>
</tr>
<tr>
<td>$T_s$</td>
<td>solidus temperature</td>
</tr>
<tr>
<td>$V_f$</td>
<td>fluid velocity</td>
</tr>
<tr>
<td>$W=V_f\Phi$</td>
<td>effective fluid flux</td>
</tr>
<tr>
<td>$x$</td>
<td>depth axis [m], positive downwards</td>
</tr>
<tr>
<td>$x_{det}$</td>
<td>local depth of detachment fault in km (Appendix 7)</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>dip of bedding (Fig. 2.11)</td>
</tr>
<tr>
<td>$\beta$</td>
<td>finite amount of extension</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>dip of normal fault (Fig. 2.11)</td>
</tr>
<tr>
<td>$\Delta t=\frac{r^2L^2}{\kappa}$</td>
<td>timestep [sec]</td>
</tr>
<tr>
<td>$\Delta x=hL$</td>
<td>depthstep [m]</td>
</tr>
<tr>
<td>$\dot{\varepsilon}$</td>
<td>strain rate (constant)</td>
</tr>
<tr>
<td>$\phi$</td>
<td>dip of detachment fault (Appendix 7)</td>
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<tr>
<td>$\Phi$</td>
<td>porosity</td>
</tr>
<tr>
<td>$\kappa=\frac{K}{\rho_sc_s}$</td>
<td>thermal diffusivity of solid</td>
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<tr>
<td>$\lambda$</td>
<td>time constant for erosion [Ma.] ($E=\lambda e^{-\lambda t}Ed$)</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>specific density of solid</td>
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<tr>
<td>$\rho_f$</td>
<td>specific density of fluid</td>
</tr>
<tr>
<td>$\rho_m$</td>
<td>specific density of magma</td>
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APPENDIX 3

THE "CONSERVATION OF HEAT" EQUATION

LOOSVELD

In modified form submitted with Chapter 4 for publication in "Tectonophysics"
THE "CONSERVATION OF ENERGY" EQUATION

We start with the one-dimensional conservation of energy equation after e.g. Carslaw & Jaeger (1959),

\[ \rho_s c_s \frac{\partial T}{\partial t} = \frac{\partial}{\partial x} \left( K \frac{\partial T}{\partial x} \right) \]  

(1)

where \( T \) is temperature, \( t \) is time, \( c_s \) is the heat capacity of the solid, \( \rho_s \) is the specific density of the solid, and \( K \) is the thermal conductivity (\( \rho_s, c_s \) and \( K \) are assumed constant in each given experiment). Transfer of heat is considered in the vertical direction, \( x \), only. A list of notations for parameters and their assigned values is given in Appendix 2.

Depending on the physical processes that need to be modelled, five factors can be added to the right hand side of equation 1. The first accounts for radiogenic heat production in the crust, \( +A \). "A" is assumed to decrease exponentially with depth or to be constant (see "Initial Conditions", Appendix 5). Because of the common enrichment of heat producing elements in fractionated granites, a different \( +A_{gr} \) is assigned to the heat production in a granite.

The second factor,

\[ -E \rho_s c_s \frac{\partial T}{\partial x} \]

accounts for erosion of the upper surface, where the reference frame is fixed relative to that surface, e.g. the top boundary condition is held at the erosion level, with \( E (=v_x) \), the upward velocity of the rock system with respect to the erosion level (negative for upward movement). In the thickening experiments, a Lagrangian reference frame was generally used, such that the solid advection factor can be omitted.

The third factor,

\[ -\rho_f c_f \Phi V_f \frac{\partial T}{\partial x} \]

is the advective component for upwards single-pass fluid flow in a frame fixed to the rock pile, with \( \Phi \) the porosity, \( V_f \) the fluid velocity (negative for upward flow), \( c_f \) the heat capacity of the fluid, and \( \rho_f \) the specific density of the fluid. \( W = V_f \Phi = \) the effective fluid flux.
A fourth factor accounts for endothermic and/or exothermic reactions. Dehydration reactions are generally endothermic; e.g. the conversion of muscovite plus quartz to sillimanite plus K-feldspar absorbed approximately 47kJ/kg, which offsets a radiogenic heat production of 2.8 $10^6$W/m³ during a period of 10 Ma. (Fyfe et al., 1978, p. 131). More generally, Walther & Orville (1982) showed that the devolatilization reactions, which take place during the metamorphism of an average pelitic rock, absorb approximately 168kJ/kg, whereas some 210kJ/kg is needed to heat one kilogram of pelitic rock from 400°C to 600°C, the assumed temperature interval for such devolatilization. Thus

$$-\frac{1.68 \times 10^5 \rho_s}{200} \frac{\partial T}{\partial t}, \text{ for } 400°C < T < 600°C,$$

accounts for the heat required for endothermic devolatilization reactions during prograde metamorphism. Assuming a constant devolatilization rate during the metamorphism, this can be simplified to $(1.68 \times 10^5 \rho_s)/(\text{time of metamorphism})$. This fourth factor is often omitted, or approximated by a low value of $A_o$.

The fifth factor

$$+L_h\rho_m \frac{\partial V}{\partial t}$$

can be added to account for the fusion heat of magmas; $L_h$ is the latent heat of fusion and $\partial V/\partial t$ the derivative of the volume fraction of melt crystallised with respect to time (Wells, 1980). According to Jaeger (1964), this factor can also be approximated by an adjustment to the temperature of the intrusion, i.e. $T_i^*=T_i+L_h/c$, or by an adjustment to the specific heat of the magma, i.e. $c_m^*=c_m+L_h/(T_i-T_s)$, with $T_i-T_s$ the range of solidification of the magma. The latter approximation has been applied in some experiments here, but mostly this factor has been omitted.

Equation 1, expanded with factors 1, 2 and 3, can now be rewritten as

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial x^2} + \frac{A}{\rho_s c_s} - (E+MW) \frac{\partial T}{\partial x}$$

with $\kappa = K/\rho_s c_s$, the thermal diffusivity and $M=(\rho_F c_F)/(\rho_s c_s)$. This equation can be solved employing the Crank-Nicolson (implicit, finite-difference, second order) algorithm

$$(T_{i,j+1} - T_{i,j})/\Delta t = (\kappa/2)(T_{i+1,j+1} - 2T_{i,j+1} + T_{i-1,j+1} + T_{i,j+1} - 2T_{i,j} + T_{i-1,j})/\Delta x^2 +$$

$$+ A/\rho_s c_s - (E+MW)(T_{i+1,j} - T_{i,j})/\Delta x$$

with $\kappa = K/\rho_s c_s$, the thermal diffusivity and $M=(\rho_F c_F)/(\rho_s c_s)$. This equation can be solved employing the Crank-Nicolson (implicit, finite-difference, second order) algorithm

$$(T_{i,j+1} - T_{i,j})/\Delta t = (\kappa/2)(T_{i+1,j+1} - 2T_{i,j+1} + T_{i-1,j+1} + T_{i,j+1} - 2T_{i,j} + T_{i-1,j})/\Delta x^2 +$$

$$+ A/\rho_s c_s - (E+MW)(T_{i+1,j} - T_{i,j})/\Delta x$$

(3)
If one defines: \( r = \frac{k \Delta t}{\Delta x^2} \), "r" becomes the stability factor to the solution of the finite difference equation. Using this implicit, second order accurate scheme, the solution will be numerically stable, independent of "r" (the accuracy, however, decreases when \( r \geq 0.5 \)); we have taken "r" continuously as 0.4. After grouping subscripts, equation 3 becomes

\[
(1+0.4)T_{i,j+1} = 0.2(T_{i+1,j+1} + T_{i-1,j+1} + T_{i,j+1}) + A\Delta T/\rho_s c_s + \\
+ \left[ 1 - 0.4 + (E + MW)\Delta T/\Delta x \right] T_{i,j} + \left[ 0.2 - (E + MW)\Delta T/\Delta x \right] T_{i+1,j}
\]

(4)

so that

\[
T_{i,j+1} = [0.2(T_{i+1,j+1} + T_{i-1,j+1} + T_{i,j+1}) + A\Delta T/\rho_s c_s + \left[ 0.2 - (E + MW)\Delta T/\Delta x \right] T_{i+1,j} + \\
+ \left[ 1 - 0.4 + (E + MW)\Delta T/\Delta x \right] T_{i,j}] / 1.4
\]

(5)

After defining the initial conditions, \( T_{i,1} \), and the boundary conditions, \( T_{1,j} \) and \( T_{i,\text{max}+1,j} \), this equation can now be solved numerically.
APPENDIX 4

NEGLECTED PARAMETERS

LOOSVELD

In modified form submitted with Chapter 4 for publication in "Tectonophysics"
NEGLECTED PARAMETERS

Inevitably, numerical experiments have to be - and should be - simplified. Factors influencing metamorphic processes are numerous, and most are poorly known for a particular metamorphic history. Far from rendering numerical experiments futile, the necessary simplifications offer general and clear relations between a particular parameter and the thermal history. Here, I neglect, for reasons presented below, (i) lateral heat transfer, (ii) the depth- and temperature-dependence of physical parameters such as the thermal conductivity, (iii) convection within magma chambers, (iv) progressive deformation during the extensional experiments, and (v) frictional heating:

i The omission of lateral heat transfer is acceptable, because the experiments simulate large-scale processes (the same anti-clockwise P-T-t paths are documented or inferred for Proterozoic fold belts making up a large part of Australia). On these scales, the lateral compared to the vertical temperature gradients can safely be assumed negligible.

ii The dependence of c, Φ, K, and ρ on temperature and pressure is negligible compared to the dependence on lithology. Here, no specific lithology has been incorporated (except crust-mantle distinction), because of the widespread occurrence of the to be modelled low-P facies metamorphism.

iii Some of the largest granites are homogeneous, e.g. the pre-1800 Ma. old Kalkadoon and Ewen Batholiths (Wyborn & Page, 1983a), hence it is reasonable to assume that within these granites convection played a major role (Marsh & Maxey, 1985; Wickham, 1987). Equally so, their source rocks must have been quite homogeneous (Wyborn et al., in press). However, most of the (post-1800 Ma.) high-level granites are heterogeneous (Wyborn et al., in press) and fractionated. As we attempt to simulate the 1600-1550 Ma. old low-P facies metamorphism, it is justifiable to approximate cooling of the horizontal (I-type) granitic slabs by conduction only.

iv Finite-strain extension will result in lower temperatures than instantaneous extension. Since even instantaneous extension cannot explain the low-P facies metamorphism in the Mount Isa Inlier, simulating progressive extension is superfluous (compare also curves A1 and A2 in Fig. A8.3). Addition of the advection factor "-v_x T/∂x", with v_x the vertical velocity due to progressive strain, on the right hand side of equation 1 (Appendix 3) is therefore superfluous. (Experiment series 2, on the other hand, does incorporate progressive thickening.)

v Frictional heating (in W/m^3) Q_f=Σij τ_ij ̇ε_ij (τ_ij is stress, and ̇ε is strain rate) is neglected here as studies by Reitan (1968a, 1968b, 1969), Graham & England (1976) and Werner (1985) show that its influence is limited. Frictional melting can possibly...
lead to the development of pseudotachylytes, and especially so in cool, dry, crystalline rocks (McKenzie & Brune, 1972; Sibson, 1975; Maddock, 1983), and even to locally inverted metamorphic zones, but only if the total plate displacement is accommodated on one or only a few faults.
APPENDIX 5

MODEL PARAMETERS

LOOSVELD

In modified form submitted with Chapter 4 for publication in "Tectonophysics"
MODEL PARAMETERS

Boundary conditions. The boundary prerequisites are made up of the temperature at the surface, fixed at 0°C, and the temperature at the base of the thermal lithosphere (100-200km), $T_{\text{max}}$, which ranges from 1200°C to 1350°C. $T_{\text{max}}$ is constant in each given experiment.

Initial conditions. The initial condition for the solution of the Crank-Nicolson implicit numerical method is either a steady-state conductive geotherm, or a tectonically perturbed one. With respect to the distribution of heat producing elements (HPE) in the upper crust, two endmembers are used in this study: HPE can either be homogeneously distributed over the upper crust or their concentration decreases exponentially with depth. The exponential distribution of heat sources with depth could be the result of progressive fractionation of magmas (Lachenbruch, 1970) and upward migration of fluids with dissolution-precipitation of the HPE (Albarede, 1975; Moorbath, 1978). This exponential distribution is probably the more realistic of the two endmembers, since it preserves the observed linear relation between surface heat flow and surface radiogenic heat production under differential erosion within a heat flow province (a two-layer configuration with the upper layer of thickness $D$ and heat production $A_0$, would also locally satisfy this linear relation, but would not survive under differential erosion; Lachenbruch, 1968; Vitorello & Pollack, 1980). $A = A_0 e^{-x/D}$, with $A_0$ the radiogenic heat production at the surface, and $D$ the characteristic length scale for the distribution of HPE, in this case for the decrease of "A" with depth. Two different characteristic lengths, $D$, are used here (10 or 15 km). Both fall within the spread of characteristic lengths calculated by Vitorello & Pollack (1980). We have further assumed an $A_0$ of 2.0, 2.5, or 3.0 $\mu$Wm$^{-3}$ (Pollack & Chapman, 1977a) and a $K$ of 2 or 3Wm$^{-1}$K$^{-1}$ (Schatz & Simmons, 1972), in order to calculate various steady-state conductive geotherms:

\[
T = \frac{A_0 D^2}{K} - \frac{A_0 D^2}{K} e^{-x/D} + \frac{T_{\text{max}} - \frac{A_0 D^2}{K}}{L} x \quad (6)
\]

with $x$ in km.

$L$, the thickness of the continental lithosphere, i.e. that part of the mantle which is stabilized against solid state convection, is poorly constrained. Above the thermal lithosphere-asthenosphere boundary, the temperature-depth profile is non-adiabatic; below it, it is adiabatic. For thermal effects on the crust, the lithospheric thickness is of importance, because the time-scale of thermal relaxation is proportional to $L^2$. The lithosphere-asthenosphere boundary is probably gradational, but is here, rather
arbitrarily, represented by an isotherm. Lithosphere thickness estimates based on heat flow studies (e.g. Pollack & Chapman, 1977a, 1977b) indicate an increase of thickness with tectonic age. Jordan (1978, 1979, 1981) and Sipkin & Jordan (1976), on the basis of anomalously short travel-times of ScS-waves under Precambrian shields (an average 4 seconds shorter than under oceans), and on geochemical evidence from kimberlites, favoured 200-400 km thick rigid mantle roots under these shields. They suggested that basalt-depletion of the upper mantle results in a chemical boundary layer with a higher viscosity, while thickening of the depleted upper mantle with time explains the large tectospheric thicknesses under Precambrian shields. Woodhouse & Dziewonski (1984) confirmed high values for "L", associating mantle structures up to 350 km depth with surface tectonic expressions (ridges, shields). Richardson et al. (1984) argue, on the basis of diamond inclusions, for a lithospheric thickness of at least 180 km beneath Archaean shields. On the other hand, the thermal time constants for the lithosphere, which are very sensitive to "L" (proportional to L^2), indicate a much thinner lithosphere, i.e. approximately 125 km (Parsons & Sclater, 1977; Sclater et al., 1980). Also, the half-life of heat production for a chondritic model Earth is approximately 2000 Ma., implying a basal heat flux in the Early to Middle Proterozoic of approximately a factor two higher than Recent, resulting in a thinner lithosphere. "L" is ranged here between 100 and 200 km.

\[(\rho c_p)\epsilon_0 T/\partial x\] is the advected heat flux, with \(E\) (or \(v_{x,0}\), the characteristic vertical velocity of the medium relative to plane \(x = 0\). This is the rate of denudation of an otherwise undeforming solid (denudation can also be tectonic, by means of extension, in which case \(E\) is depth-dependent; there are, however, no indications for such late extension in the Mount Isa Inlier). "E" depends on factors as climate, resistivity against erosion of the rock types, drainage patterns and surface height, which in turn is partly governed by the rigidity of the system. All of these make generalizations about \(E\) almost impossible. Generally, \(E\) is taken either to decay exponentially with time, or to be constant. In the case of time-independent erosion-rates, values range between 2 mm/yr, as reported by Hollister (1975, 1982) and 0.1 mm/yr (=10 km/100Ma.). In four programmes, uplift is simulated with such constant erosion rates (see Appendix 6). We have generally opted for the exponential erosion function, such that, following Carson & Kirby (1972), England & Richardson (1977) and Richardson & England (1979):

\[E = \lambda E d e^{-\lambda x}\]

with \(\lambda\) a time constant for the erosional process with unit Ma. and \(Ed\) the thickness of the finally eroded sequence. "\(\lambda\)" equals \((cF)^{-1}\), with \(F=(1-\rho_c/\rho_m)\) and \(c\) another erosional time constant in km/(Ma.km elevation). \(\rho_c\) and \(\rho_m\) are respectively the density of the
eroded material (2700 kg/m³) and that of the isostatic compensation depth (3300 kg/m³). Thickness of the crust before the period of tectonism was taken as 25 or 35 km, and erosion started at the end of the crustal thickening period. In the Mount Isa Inlier, however, isobaric cooling, as deduced from the metamorphic petrology, justifies a complete omission of denudation after crustal thickening (experiment series 2).

\[ W, \text{ the effective fluid flux, equals } V_f \Phi \text{ where } V_f \text{ is the fluid velocity and } \Phi \text{ is the porosity.} \]
Thompson (1987) argues for characteristic fluid permeation rates \( (V_f) \) of 4 mm/year \( (=4 \text{ km/Ma.}=1.3 \times 10^{-10} \text{ m/s}) \). Here, similar values are used, calculated by taking \( W \) as \( 20000 \Phi / T_{\text{end}} \) in m/s, thus simulating a constant devolatilization of a 20 km thick pelitic sequence during deformation. For constant strain rate experiments with a finite amount of thickening of 1.6, this yields values of \( W \) ranging from \( 10^{-11} \) (for \( \dot{\varepsilon} = 10^{-15.4} \text{s}^{-1} \)) to \( 2.5 \times 10^{-12.5} \) (for \( \dot{\varepsilon} = 10^{-14.3} \text{s}^{-1} \)). See also §4.3.2.1.

\[ L_h, \text{ the latent heat of fusion of magma; here taken as 335kJ/kg for tonalitic compositions (Robie et al., 1978; Wells, 1980), and 420kJ/kg for a basaltic melt (Yoder, 1976).} \]

\[ \partial V / \partial t, \text{ derivative of the volume fraction of melt crystallized with respect to time.} \]
Approximated by Wells (1980) as \( (1/(T_L-T_s))(\partial T/\partial t) \), with \( T_L \), the liquidus temperature, between 800 and 1050°C, and the solidus temperature, \( T_s \), between 700 and 900°C. Both \( T_L \) and \( T_s \) depend strongly on the H₂O-content and chemical composition. The intrusion temperature, \( T_h \), was chosen arbitrarily between \( T_L \) and \( T_s \).

Fifty depthsteps were equally spaced over the lithosphere. Using a "stability factor" \( "r" = (\kappa \Delta t / \Delta x^2) \) of 0.4, this implies a timestep, \( \Delta t \), of \( 0.4 L^2/(2500 \kappa) \). With e.g. \( L=150 \text{ km} \) and \( \kappa=10^{-6} \text{ m}^2\text{s}^{-1} \), \( \Delta t=0.11 \text{ Ma.} \).
APPENDIX 6

FLOW DIAGRAM + PROGRAMMES

LOOSVELD
FLOW DIAGRAMS + PROGRAMMES

Within all programmes, equation 5 (Appendix 3) was solved in an iterative way rather than by direct matrix solution, using compiled MacBasic™ on a Macintosh Plus and, in three cases, on a Macintosh II. Using Macintoshes limits the amount of depthsteps and makes running a slow process (up to 2 hours), but their user-friendliness, availability and extensive graphics packages counterbalance this. To keep programmes short (and compilation times in the order of a few minutes), 25 different programmes were written, each catering for different degrees of sophistication of the equation for conservation of energy and/or different geometrical configurations. A wide range of tectonic configurations is covered by these 25 programmes. Modifying the programmes to simulate variations of these configurations, or to include new parameters, is simple (no previous programming experience is required). All programmes presented here have a similar structure (Fig. A6.1).

![Flow diagram of the main elements of all "Thermo-Relax (TR)" programmes. After running a programme, Microsoft Word™ is used for tabulating the datafiles. The files can then be transferred to CricketGraph™, for selected graphical output.](image)
The calculation block plus accuracy test are constructed as follows:

```
FOR j = 2 TO jmax
    FOR i = 2 TO imax
        T(i,j) = T(i,j-1)
    NEXT i
    calculate_again:
    FOR i = 2 TO imax
        cheqTEMP(i,j) = T(i,j)
        T(i,j) = (0.4*T(i-1,j-1) + 1.2*T(i,j-1) + 0.4*T(i+1,j-1) + 0.4*T(i-1,j) + 0.4*T(i+1,j)) / 2.8
    NEXT i
    FOR i = 2 TO imax 'accuracy loop
        IF T(i,j) < cheqTEMP(i,j) - Ac OR T(i,j) > cheqTEMP(i,j) + Ac THEN GOTO calculate_again '(Ac=.01)
    NEXT i
    NEXT j
```

The complete programme library is stored on a 3.5" 2S2D (800K) floppy disc. A full list is presented in Table A6.2. Programme names consist of the project abbreviation TR (for thermo-relax), the series number (1 or 2), a capital "A" or "B" (respectively indicating constant heat production in the upper crust, or exponential decrease of heat production with depth), and a number of sub- and super-scripts. Series 1 does not involve advective transfer of heat, and both the medium and the boundary conditions are fixed to the reference frame. Series 1 is used e.g. to simulate thermal relaxation after an instantaneous thermal/tectonic perturbation when no subsidence/sedimentation or uplift/erosion plays a role. Series 2, on the other hand, does involve advective heat transfer, and can be used for situations which include progressive deformation and/or erosion. Series 2 programmes have the subscript "Eul" if a Eulerian reference frame (moving medium, fixed boundary conditions) is used, and "Lag" if a Lagrangian reference frame (fixed medium, moving boundary conditions) is used. The factor which accounts for the advection of heat, -Ep c \( \partial T/\partial x \) (Appendix 3), can be omitted when using the Lagrangian reference frame. Programmes with the Eulerian reference frame in which advection is involved need additional interpolation calculations in order to construct P-T-t trajectories for any particular rock, but can be directly used for T-t relations at any particular pressure (if the grid is fixed to the Earth's surface). A list of other subscripts and their meaning is given in Table A6.1. The superscript "exp-ero" is added when the erosion rate is not constant, but decreases exponentially with time.

A special Lagrangian reference frame was set up for some of the progressive and homogeneous thickening models: the reference frame deformed with the rock medium (subscript "defLag"). The advantage of this configuration is that rockpoints coinciding with the initial mesh stay on the mesh, such that their P-T-t history will be well
constrained, without the need for interpolation; in case of a non-deforming mesh, intersections of an arbitrary rockpoint with the mesh will be very limited, as only 50 depthsteps are used here.

**Table A6.1** A list of subscripts to programme-names and their meanings

<table>
<thead>
<tr>
<th>subscript</th>
<th>programme incorporates/simulates:</th>
</tr>
</thead>
<tbody>
<tr>
<td>d</td>
<td>crust/mantle delamination</td>
</tr>
<tr>
<td>det</td>
<td>extension by means of a simple a planar detachment fault</td>
</tr>
<tr>
<td>f</td>
<td>advection of heat by fluid movement</td>
</tr>
<tr>
<td>fusion</td>
<td>latent heat of fusion</td>
</tr>
<tr>
<td>g</td>
<td>heterogeneous heat production distribution (HPE-enriched granites)</td>
</tr>
<tr>
<td>ga</td>
<td>as &quot;g&quot;, but with fluid flow above the granitic slab</td>
</tr>
<tr>
<td>i</td>
<td>intrusion of a granite at t=0</td>
</tr>
<tr>
<td>p</td>
<td>progressive thickening</td>
</tr>
<tr>
<td>thrust</td>
<td>thickening due to thrusting in the upper crust; homogeneous below that</td>
</tr>
<tr>
<td>u</td>
<td>underplating of a basaltic/andesitic slab at the base of the crust</td>
</tr>
<tr>
<td>up-pl-ext</td>
<td>homogeneous extension below a certain depth; no deformation above it</td>
</tr>
<tr>
<td>defLag</td>
<td>homogeneous progressive thickening: mesh deforms with medium</td>
</tr>
</tbody>
</table>

**Table A6.2** A list of Thermo-Relax (TR) programmes used during this project

<table>
<thead>
<tr>
<th>TR1A</th>
<th>TR1B_{i,ga},fusion</th>
</tr>
</thead>
<tbody>
<tr>
<td>TR1B</td>
<td>TR1B_{up-pl-ext}</td>
</tr>
<tr>
<td>TR1B_{i}</td>
<td>TR1B_{det}</td>
</tr>
<tr>
<td>TR1B_{i,g}</td>
<td>TR1B_{up-pl-ext,g}</td>
</tr>
<tr>
<td>TR1B_{d}</td>
<td>TR2A_{Eul}</td>
</tr>
<tr>
<td>TR2A_{expo-ero}</td>
<td>TR2B_{Eul}</td>
</tr>
<tr>
<td>TR2A_{expo-ero,Lag}</td>
<td>TR2B_{defLag,d,f,g}</td>
</tr>
<tr>
<td>TR2B_{Lag}</td>
<td>TR2B_{defLag,p,d,f}</td>
</tr>
<tr>
<td>TR2B_{defLag,p,d}</td>
<td>TR2B_{expo-ero,Lag,thrust}</td>
</tr>
</tbody>
</table>

Now follow four programmes, the first one an example of series 1, simulating instantaneous, homogeneous, "pure shear" thinning of the mantle lithosphere by a factor $\beta$ (TR1B_{up-pl-ext,g}). The second one (TR2A_{Eul}) simulates thermal relaxation of an instantaneously stretched geotherm (homogeneous thickening) synchronous with erosion, as does the third one (series 2B_{expo-ero,Lag,thrust}) but with the Lagrangian configuration and thickening by thrusting in the upper crust and homogeneous thickening in the lower crust and mantle lithosphere. Number four simulates progressive thickening of the crust accompanied by instantaneous upwelling of asthenospheric material to the base of the (deforming) crust (TR2B_{defLag,p,d,f,g}).
Appendix 6.1

REM** TR1B_up-plate-ext. g, by RAMON LOOSVELD
REM** last modified Jan, 22nd, 1988.

REM** Calculates perturbed geotherm after homogeneous, instantaneous extension
REM** below a certain depth, simulating the "upper plate" situation of the asymmetrical
REM** extension model. Subsequently, calculates relaxation of perturbed geotherm.
REM** The concentration of radioactive elements decreases exponentially with depth.
REM** Including HPE enriched granite between i=7 and 8. Calculation scheme is
REM** implicit and iterative. Data output to 7 data files.

REM** Solves ∂T/∂t = k∂²T/∂x²+Ao/(ρ*c)*e^(-x/D) (will not be non-dimensionalised).

OPTION BASE 1

Input parameters:
INPUT "number of experiment"; No
INPUT "undeformed crustal thickness in km"; CRU
INPUT "asthenospheric temperature in C"; Tmax
INPUT "thickness of undeformed (pre-extension) lithosphere in km"; L
INPUT "amount of depthsteps"; imax
INPUT "thermal diffusivity in m²/s"; diff
INPUT "dimensionless extension factor"; B
INPUT "thickness of (undeforming) "upper plate" in km"; Up
INPUT "heat production at surface, in W/m³"; Ao
INPUT "heat production granite, in W/m³"; Agr
INPUT "thickness of radioactive scale length in km"; D
INPUT "thermal conductivity in W/mK"; K
INPUT "accuracy wanted"; Ac
INPUT "all parameters OK (y/n)"; ch$
IF ch$="n" THEN GOTO Input_parameters

REM** Hard copy parameters
LPRINT" Output data to DATA FILE: TR1B_up-plate-ext. g , exp:";No
LPRINT" Undeformed crustal thickness in km: \"; TAB(55);CRU
LPRINT" Asthenospheric temperature in C ·\";TAB(55);Tmax
LPRINT" Thickness of lithosphere in km: \";TAB(55);L
LPRINT" Amount of depthsteps: \";TAB(55);imax
LPRINT" Extension factor β, mantle lithosphere: \";TAB(55);B
LPRINT" Thickness of (undeforming) "upper plate" in km";TAB(55);Up
LPRINT" Thermal diffusivity in m²/s: \";TAB(55);diff
LPRINT" Heat production at surface in W/m³\";TAB(55);Ao
LPRINT" Heat production granite in W/m³:\";Agr
LPRINT" Thickness of radioactive scale length in km: \";TAB(55);D
LPRINT" Thermal conductivity in W/mK: \";TAB(55);K
LPRINT" Accuracy: \";TAB(55);Ac

REM** Calculate necessary amount of timesteps
h=1/imax
r=.4
k=.4*h²
jmax=INT(1/k+.5)

REM** Break up of calculations in multiple time blocks (BASIC's top limit to data file)
jc=INT(32000/(imax+1))
IF jc>(jmax+1)*2/3 THEN jd=INT((jmax+1)*2/3) ELSE jd=jc
PRINT " jd=";jd
mmax=jd*INT(((jmax+1)*2/3)/jd+1)-1
DIM T(imax+1,jd), tyd(mmax+1), cheqTEMP(imax+1,jd)
Appendix 6.1

**Initial conditions** (other thermal gradients than the one below may be specified)

FOR i = 2 TO imax
  
x = (i-1)*h*L
  T(i,1) = Ao*D^2/K - (Ao*D^2/K)*EXP(-x/D) + (Tmax - Ao*D^2/K)/L*x

NEXT i

icru = INT(CRU/(h*L) + .5)  'icru represents depth level under which lithosphere is thinned
ibase = icru + INT((imax-icru)/B+.5)  'ibase: level upwards migrated base lithosphere

FOR i = 2 TO icru
  
  T(i,1) = T(i,1)

NEXT i

FOR i = icru + 1 TO ibase - 1
  
is = B * (i-icru) + icru
  
  T(i,1) = T(is,1)

NEXT i

FOR i = ibase TO imax
  
  T(i,1) = Tmax

NEXT i

LPRINT 'Initial temperatures at t=0 (km:C):'

FOR i = 2 TO imax
  
  PRINT (i-1)*h*L; ':'; INT(T(i,1)-h5); '
  
  LPRINT (i-1)*h*L; ':'; INT(T(i,1)+.5); '

NEXT i

**Boundary conditions**

FOR j = 1 TO jd
  
  T(1,j) = 0
  
  T(imax+1,j) = Tmax

NEXT j

**Get true time in [Ma]**

FOR m = 1 TO mmax+1
  
  tyd(m) = (m-1)*4*h^2*(L*1000)^2/diff
  
  tyd(m) = INT(tyd(m)/(3600*24*365*1^6)*10+.5)/10

NEXT m

LPRINT "Amount of timesteps with r=0.4 is: " ; INT((jmax+1)*2/3)

LPRINT "Timestep is " ; tyd(10)/10; "Ma"

p = 0

 calculation:

RTI = Ao*.4*h^2*L^2*10^6/K

FOR i = 2 TO imax
  FOR j = 2 TO jd
    
    T(i,j) = T(i,j-1)
  
  NEXT i

calculate_again:

FOR i = 2 TO imax
  
  cheqTEMP(i,j) = T(i,j)
  
  x = (i-1)*h*L
  
  IF i >= 7 AND i <= 8 THEN calc_special
  
    T(i,j) = (.4*T(i-1,j-1) + 1.2*T(i,j-1) + .4*T(i+1,j-1) +
    
    .4*T(i-1,j) + .4*T(i+1,j))/2.8 + (RTI/1.4)*EXP(-x/D)

  GOTO nexti

calc_special:

T(i,j) = (.4*T(i-1,j-1) + 1.2*T(i,j-1) + .4*T(i+1,j-1) + .4*T(i-1,j) +

    .4*T(i+1,j))/2.8 + (Agr*.4*h^2*L^2*10^6/(K*1.4))

  
  nexti:

  NEXT i
FOR i=2 TO imax
    IF T(i,j)<cheqTEMP(i,j)-Ac OR T(i,j)>cheqTEMP(i,j)+Ac THEN GOTO calculate_again
NEXT i
    c=c+1
    PRINT"*";c
NEXT j
    p=p+1
    PRINT "p";p

REM ** Convert to integer
FOR j=1 TO jd
    FOR i=1 TO imax+1
        T(i,j)=INT(T(i,j)+.5)
    NEXT i
NEXT j

IF p=1 THEN print_results_I
IF p=2 THEN print_results_II
IF p=3 THEN print_results_III
IF p=4 THEN print_results_IV
IF p=5 THEN print_results_V
IF p=6 THEN print_results_VI
IF p=7 THEN print_results_VII

print_results_I: ' as relaxation is most intense immediately after perturbation, storage
    ' density should be high
OPEN "datal" FOR OUTPUT AS #1
FOR j=1 TO 200 STEP 2
    m=j
    PRINT#1,tyd(m);
    FOR i=2 TO 16
        PRINT#1,T(i,j);
    NEXT i
    FOR i=17 TO imax-3 STEP 2
        PRINT#1,T(i,j);
    NEXT i
    PRINT#1,
NEXT j
FOR j=201 TO jd STEP 50
    m=j
    PRINT#1,tyd(m);
    FOR i=2 TO 16
        PRINT#1,T(i,j);
    NEXT i
    FOR i=17 TO imax-3 STEP 2
        PRINT #1,T(i,j);
    NEXT i
    PRINT #1,
NEXT j
CLOSE #1
GOTO next_array
Appendix 6.1

print_results_II: ' lower storage density
OPEN "datall" FOR OUTPUT AS #2
FOR j=2 TO jd STEP 200
  m=(jd+j-1)
  PRINT#2,tyd(m);
  FOR i=2 TO 16
    PRINT#2,T(i,j);
  NEXT i
  FOR i=17 TO imax-3 STEP 2
    PRINT #2,T(i,j);
  NEXT i
  PRINT#2,
NEXT j
CLOSE #2
GOTO next_array

print results III to print results VII idem as print results II

next_array:
IF jd*p>=INT((jmax+1)*2/3) THEN All_done
FOR i=2 TO imax
  T(i,1)=T(i,jd)
NEXT i
GOTO calculation

All_done:
PRINT"All done"
END
Appendix 6.1

Output data to DATA FILE: TR1B\textsubscript{up-plate-ext. g}, exp: 133

Asthenospheric temperature in C: 1350
Undeformed crustal thickness in km: 35
Thickness of lithosphere in km: 100
Amount of depthsteps: 50
Extension factor $\beta$, mantle lithosphere: 4
Thickness of (undeforming) "upper plate" in km: 35
Thermal diffusivity in m$^2$/s: .0000012
Heat production at surface in W/m$^3$: .0000025
Heat production granite, in W/m$^3$: .0000041
Thickness of radioactive scale length in km: 10
Thermal conductivity in W/mK: 3
Accuracy: .01

Initial temperatures at $t=0$ (km:C):
96:1350 98:1350 100:1350

Amount of timesteps with $r=0.4$ is: 4167
Timestep is .04 Ma

Data from exp. 133, TR1B\textsubscript{up-pl-ext. g}

Fig. A6.2 TR1B\textsubscript{up-plate-ext. g}: an experiment example (graphics are manipulated by CricketGraph$^\text{TM}$). The three horizontal lines represent the andalusite/sillimanite in-temperatures for the 10, 18 and 26 km level (after Holdaway, 1971).
Appendix 6.2

REM** TR2AEu as written by RAMON LOO SVELD and INEKE DE BRUYN
REM** last modified March, 10th, 1987.

REM** Relaxation of geotherm perturbed by one instantaneous thrust repetition in the
REM** upper crust. No deformation below the thrust. Erosion after the thrust event
REM** with a constant erosion rate. Eulerian reference frame.
REM** Erosion thickness equals thickness thrust.
REM** With a constant heat production in the upper crust.
REM** Data output to 7 data files.

REM** Solves ∂T/∂t = κ∂^2T/∂x^2 + A/(ρ*c) + E∂T/∂x implicitly and iteratively
REM** (equation will not be non-dimensionalised).

OPTION BASE 1

input_parameters:
INPUT "number of experiment"; No
INPUT "asthenospheric temperature in C"; Tmax
INPUT "total depth of lithosphere in km (including thrust!!)"); L
INPUT "amount of depthsteps"; imax
INPUT "thermal diffusivity in m^2/s"; diff
INPUT "heat production in upper crust in W/m^3"; A
INPUT "Thickness of radioactive scale length in km"; D
INPUT "thermal conductivity in W/mK"; K
INPUT "thickness of part that will erode in km"; Ed
INPUT "time that the erosion will last in Ma"; Et
INPUT "accuracy"; Ac
INPUT "all parameters OK (y/n)"; ch$
IF ch$="n" THEN GOTO input_parameters

REM** Hard copy parameters
LPRINT" Output data to DATA FILE: TR2AEu exp:";No
LPRINT" Asthenospheric temperature in C: ",TAB(55);Tmax
LPRINT" Total (lithospheric) depth in km (including thrust!!): ",TAB(55);L
LPRINT" Amount of depthsteps: ",TAB(55);imax
LPRINT" Thermal diffusivity in m^2/s: ",TAB(55);diff
LPRINT" Heat production in crust in W/m^3: ",TAB(55);A
LPRINT" Thickness of radioactive scale length in km: ",TAB(55);D
LPRINT" Thermal conductivity in W/mK: ",TAB(55);K
LPRINT" Thickness of part that will erode in km: ",TAB(55);Ed
LPRINT" Time erosion of thrust will take in Ma: ",TAB(55);Et
LPRINT" Accuracy: ",TAB(55);Ac

REM** Calculate necessary amount of time steps
h=1/imax
r=.4
k=.4*h^2
jmax=INT(1/k+.5)

REM** Break up of calculations in multiple time blocks (BASIC's top limit to data file)
jc=INT(32000/(imax+1))
IF jc>(jmax+1)*2/3 THEN jd=INT((jmax+1)*2/3) ELSE jd=jc
PRINT " jd=",jd
mmax=jd*INT(((jmax+1)*2/3)/jd+1)-1

DIM T(imax+1,jd),tyd(mmax+1),depth(mmax+1),cheqTEMP(imax+1,jd)
Initial conditions (after England & Thompson, 1984)

FOR i = 1 TO imax
  x = (i - 1) * h * L
  IF Ed < D THEN sub_two ELSE sub_one
  sub_one:
    IF x >= 0 AND x <= D THEN truc_I
    IF x > D AND x <= Ed THEN crust
    IF x > Ed AND x <= (Ed + D) THEN truc_II
    IF x > (Ed + D) THEN crust_mantle
  sub_two:
    IF x >= 0 AND x <= D THEN truc_I
    IF x > Ed AND x <= (Ed + D) THEN truc_II
    IF x > (Ed + D) THEN crust_mantle
  truc_I:
    T(i, 1) = A * x * (D - x/2)/K + (Tmax - A * D^2/(2 * K))/(L - Ed) * x
    GOTO nksti
  crust:
    T(i, 1) = A * D^2/(2 * K) + (Tmax - A * D^2/(2 * K))/(L - Ed) * x
    GOTO nksti
  truc_II:
    x = x - Ed
    T(i, 1) = A * x * (D - x/2)/K + (Tmax - A * D^2/(2 * K))/(L - Ed) * x
    GOTO nksti
  crust_mantle:
    x = x - Ed - D
    T(i, 1) = A * D^2/(2 * K) + (Tmax - A * D^2/(2 * K))/(L - Ed) * x
  nksti:
NEXT i

LPRINT "initial temperatures at t=0 (km:C):"
FOR i = 2 TO imax
  PRINT (i - 1) * h * L; ":"; INT(T(i, 1) + .5); " M ;
LPRINT (i - 1) * h * L; ":"; INT(T(i, 1) + .5); " M ;
NEXT i

Boundary conditions
FOR j = 1 TO jd
  T(1, j) = 0
  T(imax + 1, j) = Tmax
NEXT j

E = Ed * 1000/(Et * 10^6 * 365 * 24 * 3600)  ' E is erosion velocity in m/s

Depth with time
REM** The thickness of the lithosphere changes with time due to erosion
REM** mn is number of Δt representing erosion time Et [Ma]
mn = INT(Et * 3600 * 24 * 365 * (10)^6 * diff/((L * 1000)^2 * 4 * h^2) + .5)
REM** ER is the amount of meters eroded during one ΔT
ER = 4 * h^2 * L^2 * 10^6 * E/diff
FOR m = 1 TO mn + 1
  depth(m) = (L * 1000 - ER * (m - 1))/1000
NEXT m
FOR m = mn + 2 TO mmax + 1
  depth(m) = depth(mn + 1)
NEXT m
Appendix 6.2

REM** Get true time in [Ma]
FOR m=1 TO mmax+1
  t(yd(m))=(m-1)*.4*hA 2*(L*1000)^2/diff
  t(yd(m))=INT(t(yd(m))/(3600*24*365*10^6)*10+.5)/10
NEXT m
LPRINT" Amount of timesteps with r=0.4 is: "; INT((jmax+1)*2/3)
LPRINT" Timestep is: "; t(yd(10))/10;" Ma"

c=0
p=0
calculation:
RTI=A*.4*h^2*L^2*10^6/K
FOR j=2 TO jd
  m=(p*jd+j-p)
  IF m>=0 AND m<mn+1 THEN GOSUB erosion ELSE GOSUB erosion_ended
  c=c+1
  PRINT"* "; c
  NEXT j
p=p+i
PRINT" p="; p

REM** Convert to integer
FOR j=1 TO jd
  FOR i=1 TO imax+1
    T(i,j)=INT(T(i,j)+.5)
  NEXT i
NEXT j
IF p=1 THEN print_results_I
IF p=2 THEN print_results_II
IF p=3 THEN print_results_III
IF p=4 THEN print_results_IV
IF p=5 THEN print_results_V
IF p=6 THEN print_results_VI
IF p=7 THEN print_results_VII

print_results_I: ' as relaxation is most intense immediately after perturbation, "storage"
  density should be high
OPEN "data1" FOR OUTPUT AS #1
FOR j=1 TO 200 STEP 2
  m=j
  PRINT#1, t(yd(m));
  FOR i=2 TO 16
    PRINT#1, T(i,j);
  NEXT i
  FOR i=17 TO imax-3 STEP 2
    PRINT #1, T(i,j);
  NEXT i
  PRINT#1,
NEXT j
FOR j=201 TO jd STEP 50
  m=j
  PRINT#1, t(yd(m));
  FOR i=2 TO 16
    PRINT#1, T(i,j);
  NEXT i
NEXT i
FOR i=17 TO imax-3 STEP 2
    PRINT #1,T(i,j);
NEXT i
PRINT#1,
NEXT j
CLOSE #1
GOTO next_array

print_results_II: 'lower "storage" density
OPEN "datalT" FOR OUTPUT AS #2
FOR j=2 TO jd STEP 200
    m=(jd+j-1)
    PRINT#2,tyd(m);
    FOR i=2 TO 16
        PRINT#2,T(i,j);
    NEXT i
    FOR 1=17 TO imax-3 STEP 2
       PRINT #2,T(i,j);
    NEXT i
    PRINT#2,
NEXT j
CLOSE #2
GOTO next_array

print results III to print results VII idem to print results II

next_array:
IF jd*p>=INT((jmax+1)*2/3) THEN All_done
FOR i=2 TO imax
    T(i,1)=T(i,jd)
NEXT i
GOTO calculation

erosion:
FOR i=2 TO imax
    T(i,j)=T(i,j-1)
NEXT i
calculate_agian:
IF Ed>D THEN sub_four ELSE sub_three
sub_three:
FOR i=2 TO imax
    cheqTEMP(i,j)=T(i,j)
    x=(i-1)*h*L
    IF x>=0 AND x<=(depth(m)-(L-Ed)-(Ed-D)) THEN truc_I
    IF x>(depth(m)-(L-Ed)-(Ed-D)) AND x<=(depth(m)-(L-Ed)) THEN crust
    IF x>(depth(m)-(L-Ed)+(D) AND x<=(depth(m)) THEN crust_mantle
    IF x>depth(m) THEN asthenosphere
    truc_I:
        T(i,j)=(.4*T(i-1,j-1)+1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+.4*T(i+1,j))/2.8+
            RTi/1.4+(.4*h*L*1000*E/diff)*(T(i+1,j-1)-T(i,j-1))/1.4
GOTO nneksti
ccrust:
    T(i,j)=(.4*T(i-1,j-1)+1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+.4*T(i+1,j))/2.8+
        (.4*h*L*1000*E/diff)*(T(i+1,j-1)-T(i,j-1))/1.4
GOTO nneksti
truc_II:
    T(i,j)=(.4*T(i-1,j-1)+1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+.4*T(i+1,j))/2.8+
        RTi/1.4+(.4*h*L*1000*E/diff)*(T(i+1,j-1)-T(i,j-1))/1.4
Appendix 6.2

GOTO nneksti

ccrust_mantle:
\[ T(i,j) = \frac{(0.4 \times T(i-1,j-1) + 1.2 \times T(i,j-1) + 0.4 \times T(i+1,j-1) + 0.4 \times T(i,j))}{2.8} + \frac{(0.4 \times h \times L \times 1000 \times E/\text{diff}) \times (T(i+1,j-1) - T(i,j-1))}{1.4} \]

GOTO nneksti

asthenosphere:
\[ T(i,j) = \text{Tmax} \]

nneksti:

NEXT i

FOR i=2 TO imax
  IF T(i,j)<cheqTEMP(i,j)-Ac OR T(i,j)>cheqTEMP(i,j)+Ac THEN GOTO calculate_again
NEXT i

RETURN

sub_foun:

FOR i=2 TO imax
  cheqTEMP(i,j)=T(i,j)
  x=(i-1)*h*L
  IF x>0 AND x<=(depth(m)-(L-Ed)) THEN truc_Ia
  IF x>(depth(m)-(L-Ed)) AND x<=(depth(m)-(L-Ed-D)) THEN truc_IIa
  IF x>(depth(m)-(L-Ed)+D) AND x<=(depth(m)) THEN ccrust_mantle_a
  IF x>depth(m) THEN asthenosphere_a

truc_Ia:
\[ T(i,j) = \frac{(0.4 \times T(i-1,j-1) + 1.2 \times T(i,j-1) + 0.4 \times T(i+1,j-1) + 0.4 \times T(i,j))}{2.8} + \frac{RTI}{1.4} + \frac{(0.4 \times h \times L \times 1000 \times E/\text{diff}) \times (T(i+1,j-1) - T(i,j-1))}{1.4} \]

GOTO nneksti

truc_IIa:
\[ T(i,j) = \frac{(0.4 \times T(i-1,j-1) + 1.2 \times T(i,j-1) + 0.4 \times T(i+1,j-1) + 0.4 \times T(i,j))}{2.8} + \frac{RTI}{1.4} + \frac{(0.4 \times h \times L \times 1000 \times E/\text{diff}) \times (T(i+1,j-1) - T(i,j-1))}{1.4} \]

GOTO nneksti

ccrust_mantle_a:
\[ T(i,j) = \frac{(0.4 \times T(i-1,j-1) + 1.2 \times T(i,j-1) + 0.4 \times T(i+1,j-1) + 0.4 \times T(i,j))}{2.8} + \frac{(0.4 \times h \times L \times 1000 \times E/\text{diff}) \times (T(i+1,j-1) - T(i,j-1))}{1.4} \]

GOTO nneksti

asthenosphere_a:
\[ T(i,j) = \text{Tmax} \]

nneksti:

NEXT i

FOR i=2 TO imax
  IF T(i,j)<cheqTEMP(i,j)-Ac OR T(i,j)>cheqTEMP(i,j)+Ac THEN GOTO calculate_again
NEXT i

RETURN

erosion_ended:

FOR i=2 TO imax
  T(i,j)=T(i,j-1)
NEXT i

calcAgain:

FOR i=2 TO imax
  cheqTEMP(i,j)=T(i,j)
  x=(i-1)*h*L
  IF x>0 AND x<=(depth(m)-(L-Ed)+D) THEN truc_II
  IF x>(depth(m)-(L-Ed)+D) AND x<=depth(m) THEN ccrust_mantle
  IF x>depth(m) THEN asthenosphere

truc_II:
\[ T(i,j) = \frac{(0.4 \times T(i-1,j-1) + 1.2 \times T(i,j-1) + 0.4 \times T(i+1,j-1) + 0.4 \times T(i,j))}{2.8} + \frac{RTI}{1.4} \]
GOTO neksti
ccc.crust_mantle:
\[ T(i,j) = \frac{0.4T(i-1,j-1) + 1.2T(i,j-1) + 0.4T(i+1,j-1) + 0.4T(i-1,j) + 0.4T(i+1,j)}{2.8} \]
GOTO neksti
asthenos:
\[ T(i,j) = T_{\text{max}} \]
neksti:
NEXT i
FOR \( i = 2 \) TO imax
   IF \( T(i,j) < \text{cheqTEMP}(i,j) - \text{Ac} \) OR \( T(i,j) > \text{cheqTEMP}(i,j) + \text{Ac} \) THEN GOTO calc_again
NEXT i
RETURN
All_done:
PRINT "All done"
END

No example and results included
Appendix 6.3

REM** TR2B<sub>expo-ero</sub><sub> Lag.thrust </sub> as written by RAMON LOOSVELD
REM** Last modified April, 11th, 1987.

REM** Program TR2B<sub>expo-ero</sub><sub> Lag.thrust </sub> calculates the initial, perturbed geotherm after an
REM** instantaneous deformation event. Deformation involves thrusting in the upper
REM** crust and homogeneous thickening below that. Subsequently, the programme
REM** will calculate the thermal relaxation of the perturbed geotherm. Erosion of the
REM** thrust commences after a time delay. Erosion velocity decreases exponentially
REM** with time.
REM** Fixed medium and moving boundary conditions (Lagrangian reference frame).
REM** The concentration of radioactive elements decreases exponentially with depth.
REM** Selected data-output to 7 data files.

REM** Solves ∂T/∂t = k∂<sup>2</sup>T/∂x<sup>2</sup> + Ao/(ρ*c) implicitly and iteratively
REM** (equation will not be non-dimensionalised).

OPTION BASE 1

input_parameters:
INPUT "number of experiment";No
INPUT "asthenospheric temperature in C"; Tmax
INPUT "total depth of undeformed lithosphere in km"; L
INPUT "depth Brittle-ductile Transition in km"; BDT
INPUT "instantaneous thickening factor"; fend
INPUT "time delay for beginning erosion in Ma"; DELAY
INPUT "amount of depthsteps"; imax
INPUT "thickness of eroding thickened crust in km"; Ed
INPUT "thermal diffusivity in m<sup>2</sup>/s"; diff
INPUT "thickness of radioactive scale length in km"; D
INPUT "heat production at the surface in W/m<sup>3</sup>"; Ao
INPUT "thermal conductivity in W/mK"; K
INPUT "erosional time scale in Ma"; labda
INPUT "accuracy"; Ac
INPUT "all parameters OK (y/n)"; ch$
IF ch$="n" THEN GOTO input_parameters

REM** Hard copy parameters
LPRINT" Output data to DATA FILE: TR2B<sub>expo-ero</sub><sub> Lag.thrust </sub> exp:"; No
LPRINT" Asthenospheric temperature [C]; " ;TAB(55); Tmax
LPRINT" Undeformed lithosphere [km]; " ;TAB(55); L
LPRINT" Amount of depthsteps: " ;TAB(55); imax
LPRINT" Depth of Brittle-Ductile Transition [km]; " ;TAB(55); BDT
LPRINT" Instantaneous thickening factor:";TAB(55); fend
LPRINT" Time delay for beginning erosion [Ma]; " ;TAB(55); DELAY
LPRINT" Thickness of eroding thickened crust [km]; " ;TAB(55); Ed
LPRINT" Thermal diffusivity [m<sup>2</sup>/s]; " ;TAB(55); diff
LPRINT" Heat production at surface [W/m<sup>3</sup>]; " ;TAB(55); Ao
LPRINT" Thickness of radioactive scale length [km]; " ;TAB(55); D
LPRINT" Thermal conductivity [W/mK]; " ;TAB(55); K
LPRINT" Erosional time scale [Ma]; " ;TAB(55); labda
LPRINT" Accuracy: " ;TAB(55); Ac

REM** Calculate necessary amount of timesteps
h=1/imax
r=.4
k=.4*h<sup>2</sup>
Jmax=INT(1/k+.5)
REM** Although gradient steepens by factor fend, lithosphere "L" does not thicken by REM** factor fend  
L=L+(fend-1)*BDT ' a temperature step will thus be created at imax

REM** Break up of calculations in multiple time blocks (BASIC's top limit to data file)  
jc=INT(32000/(imax+1))
IF jc>(jmax+1)*2/3 THEN jd=INT(((jmax+1)*2/3)) ELSE jd=jc
PRINT ";jd
mmax=jd*INT(((jmax+1)*2/3)/jd+1)-1

REM** iEd represents maximum number of \Delta x's to be eroded  
iEd=INT((Ed/(h*L)))+1

DIM T(imax+1,jd),tyd(mmax+1),ER(mmax+1),cheqTEMP(imax+1,jd)
DIM Pr(imax+1,jd),td(mmax+1),mn(iEd),Temp(imax+1,jd)

REM** Initial conditions (other thermal gradients than the one below may be specified)  
FOR i=2 TO imax  
T(i,1)=Ao*D^2/K-(Ao*D^2/K)*EXP(-x/D)+(Tmax-Ao*D^2/K)/  
(L-(fend-1)*BDT)*x
NEXT i

ithrust=1+INT(BDT*(fend-1)/(h*L)+.5) 'represents base of thrust
iBDT=ithrust+INT(BDT/(h*L)+.5) 'represents base of brittle regime (below thrust)
FOR i=2 TO ithrust  
x=(i-1)*h*L
T(i,1)=Ao*D^2/K-(Ao*D^2/K)*EXP(-x/D)+(Tmax-Ao*D^2/K)/  
(L-(fend-1)*BDT)*x
NEXT i

FOR i=ithrust+1 TO iBDT  
x=((i-1)*h*L-BDT*(fend-1))  
T(i,1)=Ao*D^2/K-(Ao*D^2/K)*EXP(-x/D)+(Tmax-Ao*D^2/K)/  
(L-(fend-1)*BDT)*x
NEXT i

FOR i=iBDT TO imax  
x=(fend*(i-1)*h*L-BDT*(fend-1)-BDT)  
T(i,1)=Ao*D^2/K-(Ao*D^2/K)*EXP(-x/D)+(Tmax-Ao*D^2/K)/  
(L-(fend-1)*BDT)*x
NEXT i

LPRINT"initial temperature at t=0 (km: C):"
FOR i=2 TO imax  
PRINT (i-1)*h*L;"\":INT(T(i, 1 )+.5);"LPRINT (i-l)*h*L;"\":INT(T(i,l)+.5);"
NEXT i

REM** Boundary conditions
FOR j=1 TO jd  
T(1,j)=0  
T(imax+1,j)=Tmax
NEXT j

REM** Get true time in [Ma]  
FOR m=1 TO mmax+1  
tyd(m)=(m-1)*.4*h*L^2*(L*1000)^2/diff  
td(m)=INT(tyd(m)/(3600*24*365*10^6)*10+.5)/10
NEXT m
LPRINT" Amount of timesteps with r=0.4 is: "; INT((jmax+1)*2/3)
LPRINT" Timestep is: ";td(10)/10;"Ma"
Appendix 6.3

REM** erosion with time
REM** ER(m) is the total amount eroded at timestep m, in meters
REM** start is the timestep at which erosion starts
start = INT((DELAY*diff*3.15*10^13)/(.4*h^2*L^2*10^6))
FOR m=2 TO start
    ER(m)=0
NEXT m
FOR m=start+1 TO mmax+1
    ER(m)=Ed*1000-Ed*1000*EXP(-tyd(m-start)/(labda*3.15*10^13))
NEXT m
REM** mn(i)=number of timesteps after which (i-1) depthsteps are eroded
FOR i=1 TO iEd
    mn(i)=start+(-diff*labda*3.15*10^13)/(.4*h*L^2*10^6)*LOG(1-(i-1)*h*L/Ed)+1
NEXT i
REM** initial pressure (simplified here, as we are only interested in crustal P-T-t paths)
FOR i=1 TO imax+1
    x=(i-1)*h*L
    Pr(i,1)=INT(x*2.7*100+.5)/1000
NEXT i
c=0
p=0
pressure:
FOR j=2 TO jd
    m=(p*jd+j-p)
    FOR i=1 TO imax+1
        x=(i-1)*h*L-ER(m)/1000
        Pr(i,j)=INT(x*2.7*100+.5)/1000
        IF Pr(i,j)<0 THEN Pr(i,j)=0
    NEXT i
    NEXT j
calculation:
RTI=Ao*.4*h^2*L^2*10^6/K
FOR j=2 TO jd
    m=(p*jd+j-p)
    FOR i=2 TO imax
        T(i,j)=T(i,j-1)
    NEXT i
    calculate_again:
    FOR i=2 TO imax
        cheqTEMP(i,j)=T(i,j)
        x=(i-1)*h*L
        IF x<=ER(m)/1000 THEN air
        IF x>ER(m)/1000 AND x<=BDT*(fend-1) THEN ttthrust
        IF x>BDT*(fend-1) THEN cccrustmantle
        air:
            T(i,j)=0
            GOTO mnksti
        ttthrust:
            T(i,j)=(.4*T(i-1,j-1) +1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+
            4*T(i+1,j))/2.8+(RTI/1.4)*EXP(-x/D)
            GOTO mnksti
        cccrustmantle:
            x=x-BDT*(fend-1)
            T(i,j)=(.4*T(i-1,j-1) +1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+
            .4*T(i+1,j))/2.8+(RTI/1.4)*EXP(-x/D)
nnnksti:
NEXT i
FOR i=2 TO imax
    IF T(i,j)<cheqTEMP(i,j)-Ac OR T(i,j)>cheqTEMP(i,j)+Ac THEN
        GOTO calculate_again
NEXT i
REM** correction block corrects for discontinuous erosion in calculation block
IF j<start+2 THEN GOTO nnnekstj
FOR i=2 TO iEnd+1
    IF ((i-1)*h*L-ER(m)/1000)>0 AND ((i-1)*h*L-ER(m)/1000)<h*L THEN
correction:
    T(i,j)=T(i,j)*((i-1)*h*L-ER(m)/1000)/(h*L)
nnekstj:
    c=c+1
    PRINT"*":c
NEXT j
p=p+1
PRINT "p=":p

REM** Convert to integer
FOR j=1 TO jd
    FOR i=1 TO imax+1
        T(i,j)=INT(T(i,j)+.5)
    NEXT i
NEXT j
IF p=1 THEN print_results_I
IF p=2 THEN print_results_II
IF p=3 THEN print_results_III
IF p=4 THEN print_results_IV
IF p=5 THEN print_results_V
IF p=6 THEN print_results_VI
IF p=7 THEN print_results_VII

print_results_I: ' as relaxation is most intense immediately after perturbation, "storage"
density should be high
OPEN "datal" FOR OUTPUT AS #1
FOR j=1 TO 200 STEP 2
    m=j
        PRINT#1,td(m);
    FOR i=5 TO 13 STEP 2
        PRINT #1,Pr(i,j);
        PRINT #1,T(i,j);
    NEXT i
    FOR i=14 TO 23
        PRINT #1,Pr(i,j);
        PRINT #1,T(i,j);
    NEXT i
    PRINT#1,;
NEXT j
FOR j=201 TO jd STEP 20
    m=j
        PRINT#1,td(m);
    FOR i=5 TO 13 STEP 2
        PRINT #1,Pr(i,j);
        PRINT #1,T(i,j);
    NEXT i
FOR i=14 TO 23
  PRINT #1,Pr(i,j);
  PRINT #1,T(i,j);
NEXT i
PRINT#1,
NEXT j
CLOSE #1
GOTO next_array

print_results_II: 'lower "storage" density
OPEN "dataII" FOR OUTPUT AS #2
FOR j=2 TO jd STEP 200
  m=(jd+j-l)
  PRINT#2,td(m);
  FOR i=5 TO 13 STEP 2
    PRINT #2,Pr(i,j);
    PRINT #2,T(i,j);
  NEXT i
  FOR i=14 TO 23
    PRINT #2,Pr(i,j);
    PRINT #2,T(i,j);
  NEXT i
  PRINT#2,
NEXT j
CLOSE #2
GOTO next_array

print_results_III to print_results_VII idem as print_results_II

next_array:
IF jd*p>=INT((jmax+l)*2/3) THEN All_done
FOR i=2 TO imax
  T(i,l)=T(i,jd)
NEXT i
GOTO pressure

All_done:
PRINT"All done"
END
Appendix 6.3

Output data to DATA FILE: TR2B_{\text{exo-ero}}^{\text{Lag,thrust exp: 68}}

- Asthenospheric temperature [°C]: 1400
- Undeformed lithosphere [km]: 165
- Amount of depthsteps: 50
- Depth of Brittle-Ductile Transition [km]: 15
- Instantaneous thickening factor: 2
- Time delay for beginning erosion [Ma]: 20
- Thickness of eroding thickened crust [km]: 35
- Thermal diffusivity [m²/s]: 0.00001
- Heat production at surface [W/m²]: 0.000025
- Thickness of radioactive scale length [km]: 15
- Thermal conductivity [W/mK]: 2
- Erosional time scale [Ma]: 200
- Accuracy: 0.01

Initial temperatures at t=0 (km; °C):
29.7:280 33:294 36.3:307 39.6:321 42.9:335 46.2:348 49.5:362 52.8:376
82.5:498 85.8:512 89.1:526 92.4:539 95.7:553 99.0:567 102.3:580 105.6:594
108.9:608 112.2:621 115.5:635 118.8:649 122.1:662 125.4:676 128.7:680
132:693 135.3:707 138.6:721 141.9:734 145.2:748 148.5:762 151.8:775
155.1:789 158.4:802 161.7:815 165:1350

Amount of timesteps with r=0.4 is: 4167
Timestep is .14 Ma

Data from "exp.68,TR2B,Lag,thrust,expo-ero"

Fig. A6.3 TR2B_{\text{Lag,thrust}}^{\text{exo-ero}}: an experiment example (graphics are manipulated by CricketGraph™).
Appendix 6.4

REM** TR2Bdefl Lag. as written by RAMON LOOSVELD
REM** Last modified Jan, 9th, 1988.
REM** Relaxation of a geotherm during and after a homogeneous progressive
REM** thickening event (constant strain rate). Upwards fluid flow during deformation
REM** Lithospheric thickening results in the detachment of the mantle lithosphere à la
REM** Houseman et al. (1981). Erosion starts immediately at the end of thickening, and
REM** erosion rate decreases exponentially with time.
REM** The concentration of radioactive elements decreases exponentially with depth:
REM** Cooled granite initially 4 km thick (at i=7 and 8).
REM** Data-output to data files.
REM** During deformation: fixed medium and fixed boundary conditions, but a
REM** deforming ("Lagrangian") grid.
REM** During erosion: fixed medium and moving boundary conditions (Lagrangian
REM** reference frame).
REM** Solves ∂T/∂t = k*∂^2T/∂x^2 + Ao/(ρ*c) + E∂T/∂x + W∂T/∂x

OPTION BASE 1

input_parameters:
INPUT "number of experiment";No
INPUT "asthenospheric temperature in C";Tmax
INPUT "initial lithospheric thickness in km";L
INPUT "initial crustal thickness in km";CRU
INPUT "amount of depthsteps";imax
INPUT "thermal diffusivity in m^2/s";diff
INPUT "thermal conductivity in W/mK";K
INPUT "heat production at the surface in W/m^3";Ao
INPUT "thickness of initial radioactive scale length in km";D
INPUT "heat production of granite in W/m^3";Agr
INPUT "exponent of strain rate (e.g. 14)";SR
INPUT "finite thickening factor (e.g. 2; doubling)";fend
INPUT "factor of thickening (e.g. 1.4) 'triggering' delamination";ft
INPUT "time that delamination takes after being 'triggered' in Ma";to
INPUT "delamination temperature at base of the crust";Tempdet
INPUT "time of hot-mode delamination in Ma";thm
INPUT "thickness of finally eroded crust in km";Ed
INPUT "erosional time scale in Ma";labda
INPUT "accuracy";Ac
INPUT "all parameters OK (y/n)";ch$
IF ch$="n" THEN GOTO input_parameters

REM** W is effective fluid flow during thickening
W=(20000*12*10^(SR))/LOG(fend)

REM** Hard copy parameters
LPRINT" Output data to DATA FILE: TR2BdefLag.p.g, exp:";No
LPRINT" Astenospheric temperature in C";";TAB(68);Tmax
LPRINT" Initial lithospheric thickness in km";";TAB(68);L
LPRINT" Initial crustal thickness in km";";TAB(68);CRU
LPRINT" Amount of depthsteps";";TAB(68);imax
LPRINT" Thermal diffusivity in m^2/s";";TAB(68);diff
LPRINT" Thermal conductivity in W/mK";";TAB(68);K
LPRINT" Heat production at surface in W/m^3";";TAB(68);Ao
LPRINT" Thickness of radioactive scale length in km";";TAB(68);D
LPRINT" Heat production of granite in W/m^3";";TAB(68);Agr
LPRINT" Exponent of strain rate";";TAB(68);SR
LPRINT" Finite thickening factor: ";TAB(68);fend
LPRINT" Effective fluid flow during thickening in m/s: ";TAB(68);W
LPRINT" Factor of thickening 'triggering' delamination: ";TAB(68);ft
LPRINT" Time that delamination takes after being 'triggered' in Ma: ";TAB(68);to
LPRINT" Delamination temperature at base of the crust in C: ";TAB(68);Tempdet
LPRINT" Time of hot-mode delamination in Ma: ";TAB(68);thm
LPRINT" Thickness of finally eroded crust in km: ";TAB(68);Ed
LPRINT" Erosional time scale in Ma: ";TAB(68);labda
LPRINT" Accuracy: ";TAB(68);Ac

REM** Calculate necessary amount of time steps

h=l/imax
r=.4
jmax=INT(1/k+.5)

REM** Break up of calculations in multiple time blocks (BASIC's top limit to data file)

jc=INT(32000/(imax+1))
IF jc>(jmax+1)*2/3 THEN jd=INT((jmax+1)*2/3) ELSE jd=jc
PRINT "jd=";jd
mmax=jd*INT(((jmax+1)*2/3)/jd+1)-1

REM** iEd represents maximum number of Δx's to be eroded
iEd=INT(Ed/(h*L)) +1

DIM T(imax+1,jd),Er(mmax+1),cheqTEMP(imax+1,jd)
DIM PR(imax+1,jd),td(mmax+1)

REM** Initial temperature conditions (other thermal gradients than the one below may
REM** be specified)
FOR i=2 TO imax
  x=(i-1)*h*L
  T(i,1)=Ao*D2/K-(Ao*D2/K)*EXP(-x/D)+(Tmax-Ao*D2/K)/L*x
NEXT i
LPRINT" Initial temperatures at t=0 (km:C):"
FOR i=2 TO imax
  PRINT (i-0.5);";INT(T(i,1)+.5);";
NEXT i

REM** tfend is time at end of thickening in sec
tfend=(LOG(fend))*10^2(SR)

REM** tdet= time of detachment (tft+to) in sec
tft=10^2(SR)*LOG(ft)
tdet=tf+to*3.15*10^13
idet=INT(CRU/(h*L)+.5)

REM** Get true time in seconds
td(1)=INT((.4*h^2*L^2*10^6/diff)*10+.5)/10
FOR m=2 TO mmax+1
  td(m)=td(m-1)+INT((.4*h^2*(L*1000*EXP(10^(-SR)*td(m-1)))^2/diff)*10+.5)/10
  IF tdet-td(m)<=0 THEN Ajtdet
NEXT m
Ajtdet
jtdet=m
LPRINT" mtdet and tdet are: ";m; td(m)/(3.15*10^13);" Ma"
FOR m=jtdet+1 TO mmax+1
    td(m)=td(m-1)+INT(((4*h^2*(L*1000*EXP(10^(-SR)*td(m-1)))^2/diff)*10+.5)/10
    IF td(m)<0 THEN Bjthm
NEXT m

Bjthm:
    mthm=m
    thm=td(m)/(3.15*10^13);"Ma"
FOR m=jthm+1 TO mmax+1
    td(m)=td(m-1)+INT(((4*h^2*(L*1000*EXP(10^(-SR)*td(m-1)))^2/diff)*10+.5)/10
    IF tfend-td(m)<=0 THEN Cjtfend
NEXT m

Cjtfend:
    jtfend=m
    tfend=td(m)/(3.15*10^13);"Ma"
FOR m=jtfend+1 TO mmax+1
    td(m)=td(jtfend)+INT(((m-jtfend)*4*h^2*(L*1000*fend)^2/diff)*10+.5)/10
NEXT m

Amount of calculated timesteps with r=0.4 is: INT((jmax+1)*2/3)

Initial timestep is: (r*h^2*L/2*10^6)/(diff*t3.15*10^13);"Ma"

** Boundary conditions
FOR j=1 TO jd
    T(1,j)=0
    T(imax+1,j)=Tmax
NEXT j

** Initial pressure
xidet=(idet-1)*h*L
FOR i=1 TO idet
    x=(i-1)*h*L
    PR(i,1)=INT(x*2.7*100+.5)/1000
NEXT i
FOR i=idet+1 TO imax+1
    x=(i-1)*h*L
    PR(i,1)=PR(idet,1)+INT((x-xidet)*3.2*100+.5)/1000
NEXT i

** pressure up to end of thickening
pressure_1_2_3:
FOR j=2 TO jtfend
    m=j
    xidet=CRU*EXP(td(m)*10^(-SR))
    FOR i=1 TO idet
        x=(i-1)*h*L*EXP(td(m)*10^(-SR))
        PR(i,j)=INT((x*2.7*100+.5)/1000
    NEXT i
FOR i=idet+1 TO imax+1
    x=(i-1)*h*L*EXP(td(m)*10^(-SR))
    PR(i,j)=PR(idet,j)+INT((x-xidet)*3.2*100+.5)/1000
NEXT i
NEXT j
c=0
p=0
REM** calculation temperatures until detachment
calculation_1:
RTI=\text{Ao} \cdot 4 \cdot \text{h}^2 \cdot \text{L}^2 \cdot 10^6 / 6
FOR j=2 TO jtdet
m=j
f=\text{EXP}((\text{td}(m) \cdot 10\cdot(-\text{SR}))
\text{dx}=\text{f} \cdot \text{h} \cdot \text{L} \cdot 1000
\text{dt}=4 \cdot \text{h}^2 \cdot \text{L}^2 \cdot 10^6 / \text{diff}
\text{Wef} = \text{W} \cdot \text{dt} / (\text{dx} \cdot 1.4)
FOR i=2 TO imax
T(i,j)=T(i,j-1)
NEXT i
calculate_again_1:
FOR i=2 TO imax
\text{cheqTEMP}(i,j)=T(i,j)
\text{x}=(i-1) \cdot \text{h} \cdot \text{L}
\text{IF} i=7 \text{ AND } i<=8 \text{ THEN} \text{calc_special}
\text{T}(i,j)=0.4 \cdot \text{T}(i-1,j-1)+1.2 \cdot \text{T}(i,j-1)+0.4 \cdot \text{T}(i+1,j-1)+0.4 \cdot \text{T}(i-1,j)+
0.4 \cdot \text{T}(i+1,j)/2.8+ (\text{RTI} \cdot \text{f}^2 / 1.4) \cdot \text{EXP}(-x/D)+\text{Wef} \cdot (\text{T}(i+1,j-1)-\text{T}(i,j-1))
\text{GOTO nexti}
\text{calc_special:}
\text{T}(i,j)=0.4 \cdot \text{T}(i-1,j-1)+1.2 \cdot \text{T}(i,j-1)+0.4 \cdot \text{T}(i+1,j-1)+
0.4 \cdot \text{T}(i-1,j)+0.4 \cdot \text{T}(i+1,j))/2.8+ (\text{Ag}\cdot 4 \cdot \text{h}^2 \cdot \text{L}^2 \cdot 10^6 \cdot \text{f}^2 / (\text{K} \cdot 1.4))+
\text{Wef} \cdot (\text{T}(i+1,j-1)-\text{T}(i,j-1))
\text{nexti:}
\text{NEXT i}
\text{FOR} i=2 \text{ TO imax}
\text{IF} \text{T}(i,j)<\text{cheqTEMP}(i,j)-\text{Ac} \text{ OR } \text{T}(i,j)>\text{cheqTEMP}(i,j)+\text{Ac} \text{ THEN}
calculate_again_1
\text{NEXT i}
c=c+1
\text{PRINT"*";c}
\text{NEXT j}
REM** calculation temperatures until end of hot mode delamination
calculation_2:
FOR j=jtdet+1 TO jthm
m=j
f=\text{EXP}((\text{td}(m) \cdot 10\cdot(-\text{SR}))
\text{dx}=\text{f} \cdot \text{h} \cdot \text{L} \cdot 1000
\text{dt}=4 \cdot \text{h}^2 \cdot \text{L}^2 \cdot 10^6 / \text{diff}
\text{Wef} = \text{W} \cdot \text{dt} / (\text{dx} \cdot 1.4)
FOR i=2 TO idet
T(i,j)=Tempdet
NEXT i
\text{FOR} i=idet+1 \text{ TO imax}
T(i,j)=Tempdet
\text{NEXT i}
calculate_again_2:
\text{FOR} i=2 \text{ TO idet}
\text{cheqTEMP}(i,j)=T(i,j)
\text{x}=(i-1) \cdot \text{h} \cdot \text{L}
\text{IF} i>=7 \text{ AND } i<=8 \text{ THEN} \text{calc_special_2}
\text{T}(i,j)=0.4 \cdot \text{T}(i-1,j-1)+1.2 \cdot \text{T}(i,j-1)+0.4 \cdot \text{T}(i+1,j-1)+0.4 \cdot \text{T}(i-1,j)+
0.4 \cdot \text{T}(i+1,j)/2.8+ (\text{RTI} \cdot \text{f}^2 / 1.4) \cdot \text{EXP}(-x/D)+\text{Wef} \cdot (\text{T}(i+1,j-1)-\text{T}(i,j-1))
\text{GOTO nexti_2}
calc_special_2:
T(i,j) = (.4*T(i-1,j-1) + 1.2*T(i,j-1) + .4*T(i+1,j-1) + .4*T(i-1,j) + .4*T(i+1,j))/2.8 + (Agr*.4*h^2*L^2*10^6*f^2/(K*1.4)) + Wef*(T(i+1,j-1)-T(i,j-1))

nexti_2:
NEXT i
FOR i=idet+1 TO imax
  cheqTEMP(i,j) = T(i,j)
  T(i,j) = Tempdet
NEXT i
FOR i=2 TO imax
  IF T(i,j) < cheqTEMP(i,j) - Ac OR T(i,j) > cheqTEMP(i,j) + Ac THEN
    GOTO calculate_again_2
  NEXT i
  c = c+1
  PRINT "*"; c
NEXT j

REM** calculation temperatures until end of thickening
calculation_3:
FOR j=jthm+1 TO jtfend
  m = j
  f = EXP(td(m)*10^(-SR))
  dx = f*h*L*1000
  dt = .4*h^2*L^2*10^6/diff
  Wef = W*dt/(dx*1.4)
  FOR i=2 TO imax
    T(i,j) = T(i,j-1)
    T(i,j) = T(i,j-1)
    IF i>=7 AND i<=8 THEN
calc_special_3
    T(i,j) = (.4*T(i-1,j-1) + 1.2*T(i,j-1) + .4*T(i+1,j-1) + .4*T(i-1,j) + .4*T(i+1,j))/2.8 + (RTI*f^2/1.4)*EXP(-x/D) + Wef*(T(i+1,j-1)-T(i,j-1))
    GOTO nexti_3
    calc_special_3:
    T(i,j) = (.4*T(i-1,j-1) + 1.2*T(i,j-1) + .4*T(i+1,j-1) + .4*T(i-1,j) + .4*T(i+1,j))/2.8 + (Agr*.4*h^2*L^2*10^6*f^2/(K*1.4)) + Wef*(T(i+1,j-1)-T(i,j-1))
  NEXT j
  FOR i=2 TO imax
    IF T(i,j) < cheqTEMP(i,j) - Ac OR T(i,j) > cheqTEMP(i,j) + Ac THEN
      GOTO calculate_again_3
    NEXT i
    c = c+1
    PRINT "*"; c
NEXT j
p = p+1
PRINT "p="; p
GOTO convert_to_integer_1
erosion_with_time:
REM** ER(m) is the total amount eroded at timestep m, in kilometres
FOR m=2 TO jtfend
    Er(m)=0
NEXT m
FOR m=jtfend+1 TO mmax+1
    Er(m)=Ed-Ed*EXP(-(td(m)-td(jtfend))/(labda*3.15*10^13))
NEXT m

pressure_4:
FOR j=2 TO jd
    m=(jtfend+(p-1)*jd+j-p)
    xidet=fend*(idet-1)*h*L-Er(m)
    FOR i=1 TO idet
        x=fend*(i-1)*h*L-Er(m)
        PR(i,j)=INT(x*2.7*100+.5)/1000
        IF PR(i,j)<0 THEN PR(i,j)=0
    NEXT i
    FOR i=idet+1 TO imax+1
        x=fend*(i-1)*h*L-Er(m)
        PR(i,j)=PR(idet,j)+INT((x-xidet)*3.2*100+.5)/1000
        IF PR(i,j)<0 THEN PR(i,j)=0
    NEXT i
NEXT j

REM** calculation temperatures during erosion times in a Lagrangian reference frame
calculation_4:
FOR j=2 TO jd
    m=jtfend+(p-1)*jd+j-p
    FOR i=2 TO imax
        T(i,j)=T(i-1,j-1)
    NEXT i
    calculate_again_4:
    FOR i=2 TO imax
        cheqTEMP(i,j)=T(i,j)
        x=fend*(i-1)*h*L-Er(m)
        IF x>0 THEN rock
            T(i,j)=0
        GOTO nexti_4
    rock:
        IF x<h*L*fend THEN correction
            x=(i-1)*h*L
            IF i>=7 AND i<=8 THEN calcspecial_4
                T(i,j)=((.4*T(i-1,j-1) +1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+.4*T(i+1,j))/2.8+(RTI*fend*A2/1.4)*EXP(-x/D))
        GOTO nexti_4
    calcspecial_4:
        T(i,j)=((.4*T(i-1,j-1) +1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+.4*T(i+1,j))/2.8+(Agr*.4*h*A2*L^2*10^6*f^2/(K*1.4))
        GOTO nexti_4
    correction:
        x=(i-1)*h*L
        IF i>=7 AND i<=8 THEN calcspecial_5
            T(i,j)=((i-1)*fend*h*L*Er(m))/(fend*h*L)*((.4*T(i-1,j-1) +1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+.4*T(i+1,j))/2.8+
            (RTI*fend*A2/1.4)*EXP(-x/D))
        GOTO nexti_4
Appendix 6.4

\[ T(i,j) = (0.4T(i-1,j-1) + 1.2T(i,j-1) + 0.4T(i+1,j-1) + 0.4T(i-1,j) + 0.4T(i+1,j))/2.8 + (\text{Agr} \times 0.4 \times h^2 \times L^2 \times 10^6 \times f^2 / (K \times 1.4)) \]

nexti_4:
FOR i = 2 TO imax
    IF T(i,j) < cheqTEMP(i,j) - Ac OR T(i,j) > cheqTEMP(i,j) + Ac THEN GOTO calculate_again_4
NEXT i

c = c + 1
PRINT "*"; c

NEXT j

p = p + 1
PRINT "p = "; p
GOTO convert_to_integer_2

convert_to_integer_1:
FOR j = 1 TO jtfend
    FOR i = 1 TO imax + 1
        T(i,j) = INT(T(i,j) + 0.5)
    NEXT i
    NEXT j
GOTO print_results_1

convert_to_integer_2:
FOR j = 1 TO jd
    FOR i = 1 TO imax + 1
        T(i,j) = INT(T(i,j) + 0.5)
    NEXT i
    NEXT j

IF p = 2 THEN print_results_II
IF p = 3 THEN print_results_III
IF p = 4 THEN print_results_IV
IF p = 5 THEN print_results_V
IF p = 6 THEN print_results_VI
IF p = 7 THEN print_results_VII
IF p = 8 THEN print_results_VIII
IF p = 9 THEN print_results_IX

print_results_I: ' as relaxation is most intense immediately after perturbation, "storage" density should be high
OPEN "data1" FOR OUTPUT AS #1
FOR j = 1 TO jtfend STEP 4
    m = j
    PRINT #1, INT((0.5 + 10 * td(m)) / (3.15 * 10^13)) / 10;
    FOR i = 2 TO 10 STEP 2
        PRINT #1, PR(i,j);
        PRINT #1, T(i,j);
    NEXT i
    FOR i = 11 TO 20
        PRINT #1, PR(i,j);
        PRINT #1, T(i,j);
    NEXT i
    PRINT #1,
NEXXT j
CLOSE #1
FOR i=2 TO imax
    T(i,1)=T(i,jtfend)
NEXT i
GOTO erosion_with_time

print_results_II: ' lower "storage" density
OPEN "datalII" FOR OUTPUT AS #2
FOR j=2 TO jd STEP 10
    m=(jtfend+j-l)
    PRINT#2,INT(.5+10*td(m)/(3.15*10^13))/10;
    FOR i=2 TO 10 STEP 2
        PRINT #2,PR(i,j);
        PRINT #2,T(i,j);
    NEXT i
    FOR i=11 TO 20
        PRINT #2,PR(i,j);
        PRINT #2,T(i,j);
    NEXT i
    PRINT#2,
NEXT j
CLOSE #2
GOTO next_array

print_results_III to print_results_IX idem as print_results_II

next_array:
    IF td(m) > 1.575*10^16 THEN All_done
    IF jd*p>=INT((jmax+l)*2/3) THEN All_done
    FOR i=2 TO imax
        T(i,1)=T(i,jd)
    NEXT i
    GOTO pressure_4

All_done:
    PRINT "All done"
END
Appendix 6.4

Output data to DATA FILE: TR2B_{e\$p-o\$e} exp: 126

Asthenospheric temperature in C: 1350
Initial lithospheric thickness in km: 150
Initial crustal thickness in km: 25
Amount of depthsteps: 50
Thermal diffusivity in m^2/s: 0.000012
Thermal conductivity in W/mK: 3
Heat production at surface in W/m^3: 0.000025
Thickness of radioactive scale length in km: 10
Heat production of granite in W/m^3: 0.000041
Exponent of strain rate: 14.82
Finite thickening factor: 1.6
Effective fluid flow during thickening in 10^-9 m/s: 0.0076
Factor of thickening 'triggering' delamination: 1
Time that delamination takes after being 'triggered' in Ma: 0
Delamination temperature at base of the crust in C: 1300
Time of hot-mode delamination in Ma: 10
Thickness of finally eroded crust in km: 0
Erosional time scale in Ma: 500
Accuracy: 0.01

Initial temperatures at t=0 (km:C):
144:1302 147:1326 150:1350

tmidet and tdet are: 2.0848273 Ma
mtthm and thm are: 153 10.02672 Ma
mtfend and tfend are: 154 10.13684 Ma

Amount of calculated timesteps with r=0.4 is: 4167
Initial timestep is 4.232804E-02 Ma
Timestep after thickening is: .1083598 Ma

Data from exp. 126, TR2B_{e\$p-o\$e} Lag,p,d,f,g,expo-ero

Fig. A6.4 TR2B_{e\$p-o\$e} Lag,p,d,f,g,expo-ero: an experiment example (graphics are manipulated by CricketGraph™).
Programmes were checked (a) by solving "crude-grid" problems using the EXCEL™ spreadsheet on a Mackintosh Plus, (b) by comparing results with those of more simple explicit schemes, and (c) by imitating experiments from published numerical and analytical studies (respectively England & Thompson, 1984, and Voorhoeve & Houseman, 1988) and comparing results. All three methods gave excellent correlations.

As an example, an analytical experiment by Voorhoeve & Houseman (1988) is imitated. In this experiment, lithospheric thinning by the asymmetric detachment model (Wernicke, 1985) and subsequent thermal relaxation are simulated. The detachment fault is planar and transects the entire lithosphere. By defining the heave and the dip of the fault, the amount and style of extension is defined. Following McKenzie (1978), Voorhoeve & Houseman (1988) argued that as extension occurs in a time short compared to the thermal time constant of the lithosphere, it is justified to approximate the

**Fig. A7.1** Geometrical elements of the asymmetrical extension model, with a planar, throughgoing detachment fault cutting through the entire lithosphere; fault dip \( \phi \) and heave (horizontal displacement) \( e \). From Voorhoeve & Houseman (1988).
extensional process as being instantaneous. Thus, the problem is reduced to the thermal relaxation of an instantaneously deformed geotherm. By neglecting lateral heat transfer, the thermal history of a particular rock depends on the initial and boundary conditions, the conductivity, thermal diffusivity, distribution of radioactive elements, the heave and dip of the detachment fault, and the lateral location, which determines the local depth to the detachment (and step in the initial geotherm). With no radiogenic heat production, the conservation of energy equation
\[ \rho C_s \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial x^2} \]
has a time-dependent solution (with the initial condition of the "stepped" geotherm)

\[ T = T_{\text{max}} x' + T_{\text{max}} \sum_{n=1}^{\infty} b_n e^{-n^2 \tau t'} \sin(n \pi x') \]

(in non-dimensionalized form) with \( t' = \frac{ktd^2}{L^2} \) and \( x' = x/L \) and

\[ b_n = \frac{2}{n\pi} \left\{ (1-c) \cos(n\pi a) + \frac{1}{n\pi} \sin(n\pi c) \right\} \]

with \( aL \) the thickness of that part of the lithosphere which lies above the detachment, and \( c = (T_{\text{max}} - T_{\text{dif}})/T_{\text{max}} \) and \( T_{\text{dif}} \) is the temperature difference over the step in the initial geotherm (see Fig. A7.1).

The graphical solution to the case with a heave of 100 km on a detachment which dips 12° and the detachment at 20 km depth is presented in Fig. A7.2. The transparent overlay is the solution to the same problem but solved numerically with TRIBdet. Programme TRIBdet is, in slightly condensed form, given at pages 201-204.
Output data to DATA FILE TR1Bdet, exp: 87

- Asthenospheric temperature in °C: 1333
- Undeformed lithospheric thickness in km: 120
- Amount of depthsteps: 50
- Thermal diffusivity in m²/s: 0.0000008
- Heat production at surface in W/m³: 0
- Thickness of radioactive scale length in km: -
- Thermal conductivity in W/Km: 3
- Local depth of detachment in km: 20
- Dip of detachment in degrees: 12
- Horizontal extension (heave) in km: 100
- Accuracy: .01

"Initial temperatures at t=0 (km:C):"

Amount of timesteps with r=0.4 is: 4167
Timestep is .09 Ma

Fig. A7.2  Comparison of results of solutions obtained by (A) analytical (Fourier) means (from Voorhoeve & Houseman, 1988) and (B) numerical modelling (this study). See text for further explanation.
Appendix 7

REM** TR1Bdet as written by RAMON LOOSVELD
REM** Last modified April, 2nd, 1987.

REM** Calculates perturbed geotherm after instantaneous detachment faulting. Then REM** calculates thermal relaxation of perturbed geotherm. Geological reference:
REM** Voorhoeve & Houseman (1988). Basin Research, 1. The concentration of
REM** radioactive elements decreases exponentially with depth.
REM** Solves ∂T/∂t=κ∂²T/∂x² + (A₀/(c*p))*e^((-x^D)/) in an implicit, iterative way
REM** (will not be non-dimensionalised). Data output to 7 data files.

OPTION BASE 1

input_parameters:
INPUT "number of experiment";No
INPUT "asthenospheric temperature in C"; Tmax
INPUT "undeformed lithospheric thickness in km"; L
INPUT "amount of depthsteps"; imax
INPUT "thermal diffusivity in m²/s"; diff
INPUT "heat production at surface in W/m³"; Ao
INPUT "thickness of radioactive scale length in km"; D
INPUT "thermal conductivity in W/Km"; K
INPUT "local depth of detachment in km"; xdet
INPUT "dip of detachment in degrees"; alpha
INPUT "horizontal extension (heave) in km"; heave
INPUT "accuracy wanted"; Ac
INPUT "all parameters OK (y/n)"; ch$
s
REM** Hard copy parameters
LPRINT" Output data to DATA FILE TRIBdet, exp: ";No
LPRINT" Asthenospheric temperature in C: ";TAB(55);Tmax
LPRINT" Undeformed lithospheric thickness in km: ";TAB(55);L
LPRINT" Amount of depthsteps: ";TAB(55);imax
LPRINT" Thermal diffusivity in m²/s: ";TAB(55);diff
LPRINT" Heat production at surface in W/m³: ";TAB(55);Ao
LPRINT" Thermal conductivity in W/Km: ";TAB(55);K
LPRINT" Thickness of radioactive scale length in km: ";TAB(55);D
LPRINT" Local depth of detachment in km: ";TAB(55);xdet
LPRINT" Dip of detachment in degrees: ";TAB(55);alpha
LPRINT" Horizontal extension (heave) in km: ";TAB(55);heave
LPRINT" Accuracy: ";TAB(55);Ac

throw=heave*TAN(alpha/57)

REM** Calculate necessary amount of time steps
h=1/imax
r=.4
k=.4*h²
jmax=INT(1/k+.5)

REM** Break up of calculations in multiple time blocks (BASIC's top limit to data file)
jc=INT(32000/(imax+1))
IF jc>(imax+1)*2/3 THEN jd=INT((jmax+1)*2/3) ELSE jd=jc
PRINT " jd=";jd
mmax=jd*INT(((jmax+1)*2/3)/jd+1)-1

DIM T(imax+1,jd), tyd(mmax+1), cheqTEMP(imax+1,jd)
Appendix 7

REM** Initial conditions (other thermal gradients than the one below may be specified)
PRINT" initial temperature at t=0 (km:C):"
idet=INT(xdet*imax/L+.5)
iad=imax-INT(throw*imax/L+.5)
FOR i=2 TO idet 
x=(i-1)*h*L*1000
T(i,1)=A0*D^2/K-(A0*D^2/K)*EXP(-x/D)+(Tmax-A0*D^2/K)/L*x
NEXT i
FOR i=idet+1 TO iad 
x=throw+(i-1)*h*L*1000
T(i,1)=A0*D^2/K-(A0*D^2/K)*EXP(-x/D)+(Tmax-A0*D^2/K)/L*x
NEXT i
FOR i=iad+1 TO imax 
T(i,1)=Tmax
NEXT i
FOR i=2 TO imax 
PRINT (i-1)*h*L;"km: ";INT(T(i,1)+.5);"C ";
LPRINT (i-1)*h*L;"km: ";INT(T(i,1)+.5);"C ";
NEXT i

REM** Boundary conditions
FOR j=1 TO jd 
T(1,j)=0
T(imax+1,j)=Tmax
NEXT j

REM** Get true time in [Ma]
FOR m=1 TO mmax+1
  tyd(m)=(m-1)*.4*h*A2*(L*A2*10^6/diff
  tyd(m)=INT(tyd(m)/(3600*24*365*10^6)*10+.5)/10
NEXT m
LPRINT" Amount of timesteps with r=0.4 is: "; INT((jmax+1)*2/3)
LPRINT" Timestep is "; tyd(10)/10; "Ma"

p=0
c=0
calculation:
RTI=A0*.4*h*A2*L*10^6/K
FOR j=2 TO jd 
FOR i=2 TO imax 
T(i,j)=T(i,j-1)
NEXT i
calculate_again:
FOR i=2 TO imax 
  cheqTEMP(i,j)=T(i,j)
x=(i-1)*h*L*1000 
  T(i,j)=(.4*T(i-1,j-1)+1.2*T(i,j-1)+.4*T(i+1,j-1)+.4*T(i-1,j)+.4*T(i+1,j))/2.8 +(RTI/1.4)*EXP(-x/(D*1000))
NEXT i
FOR i=2 TO imax 
  IF T(i,j)<cheqTEMP(i,j)-Ac OR T(i,j)>cheqTEMP(i,j)+Ac THEN GOTO calculate_again
NEXT i
  c=c+1
PRINT" ";c
NEXT j
p=p+1
PRINT "p":p
REM** Convert to integer
FOR j=1 TO jd
   FOR i=1 TO imax+1
      T(i,j)=INT(T(i,j)+.5)
   NEXT i
NEXT j

IF p=1 THEN print_results_I
IF p=2 THEN print_results_II
IF p=3 THEN print_results_III
IF p=4 THEN print_results_IV
IF p=5 THEN print_results_V
IF p=6 THEN print_results_VI
IF p=7 THEN print_results_VII

print_results_I: ' as relaxation is most intense immediately after perturbation, "storage'
' density should be high

OPEN "datal" FOR OUTPUT AS #1
FOR j=1 TO 200 STEP 2
   m=j
   PRINT#1,tyd(m);
   FOR i=2 TO 16
      PRINT#1,T(i,j);
   NEXT i
   FOR i=17 TO imax-3 STEP 2
      PRINT#1,T(i,j);
   NEXT i
   PRINT#1,
NEXT j
FOR j=201 TO jd STEP 50
   m=j
   PRINT#1,tyd(m);
   FOR i=2 TO 16
      PRINT#1,T(i,j);
   NEXT i
   FOR i=17 TO imax-3 STEP 2
      PRINT#1,T(i,j);
   NEXT i
   PRINT#1,
NEXT j
CLOSE #1
GOTO next_array

print_results_II: ' lower storage density
OPEN "dataII" FOR OUTPUT AS #2
FOR j=2 TO jd STEP 200
   m=(jd+j-1)
   PRINT#2,tyd(m);
   FOR i=2 TO 16
      PRINT#2,T(i,j);
   NEXT i
   FOR i=17 TO imax-3 STEP 2
      PRINT#2,T(i,j);
   NEXT i
   PRINT#2,
NEXT j
CLOSE #2
GOTO next_array

print_results_III to print_results_VII idem as print_results_II
next_array:
IF jd*p>=INT((jmax+1)*2/3) THEN All_done

FOR i=2 TO imax
   T(i,1)=T(i,jd)
NEXT i
GOTO calculation

All_done:
PRINT"All done"
END
APPENDIX 8

A MODEL FOR LOW-PRESSURE FACIES METAMORPHISM DURING CRUSTAL THICKENING

LOOSVELD
ETHERIDGE

Submitted to "Nature"
There is considerable discussion (Vissers, pers. comm., 1988; Zwart, 1967; Wickham & Orinburg, 1987; 1985; Verhoef et al., 1984) as to whether low-P facies metamorphism in the Pyrenees is synchronous with a phase of regional extension or shortening.
INTRODUCTION

In a recent influential article, Wickham & Oxburgh (1985) proposed that the very steep metamorphic gradients (80-100°C/km) deduced from the Trois Seigneurs Massif, eastern Pyrenees, resulted from continental extension (Fig. A8.1A). Lithospheric extension was considered to have led to an attenuated series of isograds, anomalously high base crustal heat flow, crustal and upper mantle melting, and low-P facies metamorphism. This model for low-P facies metamorphism was further developed by Sandiford & Powell (1986), and is supported by measurements of high heat flow in present-day continental rifts (Edwards et al., 1978; Lachenbruch, 1979; Lachenbruch & Sass, 1977; Morgan, 1982, 1983).

Low-P facies metamorphism, characterized by andalusite-sillimanite facies, is widespread among fold belts of all ages, but particularly those of the Precambrian, where it is the rule. In most (if not all) of these fold belts, however, the low-P facies assemblages are synchronous with compressional deformation, evidenced by thrusting and/or tight, upright folding. Heating therefore accompanied thickening rather than thinning of the crust. In some of these terranes (e.g. the Mount Isa Inlier and Broken Hill Province, see below) there is convincing evidence that prograde low-P facies metamorphism postdated the closest period of continental extension by as much as 100 Ma. In such cases, the thermal anomaly resulting from extension would have largely decayed, and cannot explain the syn-compressional low-P facies metamorphism, let alone the prograde (+ΔT) character of it.

In this study, therefore, we propose an alternative model for low-P facies metamorphism that results directly from thickening rather than thinning of the lithosphere. There is no question that extension can give rise to crustal low-P facies metamorphism, as demonstrated by the Tertiary metamorphism associated with many of the metamorphic core complexes of the southwestern USA (Lee et al., in press). However, extensional models do have a limited potential to explain the widespread low-P facies metamorphism associated with crustal shortening in fold belts. We begin by exploring the limitations of the extensional model in terms of the magnitude and duration of the heating event. We then outline the alternative model (Fig. A8.1B), and provide two examples whose chronology, metamorphic history and tectonic setting are reasonably well constrained. Two other recently published explanations for low-P facies metamorphism in terms of mantle and crustal magmatism (Fig. A8.1C, D) are briefly discussed.
Fig. A8.1 Schematic temperature-depth diagrams showing the steady-state, conductive geotherm (left), and four perturbed geotherms. Thermal relaxation of all four perturbations leads to higher than normal T/P ratios in the middle crust. A: extension of (parts of) the lithosphere, with instantaneous, homogeneous thinning of the entire lithosphere; B: substitution of heat-conducting mantle lithosphere by heat-advecting asthenosphere during crustal thickening (model proposed here to explain low-P facies metamorphism); C: emplacement of basaltic-andesitic underplate at the base of the crust; D: emplacement of felsic magmas into the middle crust. The amount of instantaneously added heat to the lithosphere is represented by the shaded areas. M: Moho. LA: Lithosphere-asthenosphere boundary.

LIMITATIONS OF THE EXTENSIONAL MODEL

The ability of extensional models to explain low-P facies metamorphism is constrained in two ways. First, the amount of heat which is added to the crust during and after extension is limited. As the steady-state conductive geotherm is generally convex towards the temperature-axis (Fig. A8.2), instantaneous, homogeneous extension by a factor $\beta$ will not result in the multiplication of temperature on any given level by the same factor $\beta$. For example, if a steady-state geotherm, defined by $T_x = \frac{A_0 D^2 (1-e^{-x/D})^2 + q^* x}{K}$ (with $x =$ depth; reasonable parameters are: $A_0 = 2.5 \mu W/m^2; D = 10$ km; the heat flow at the base of the lithosphere, $q^* = 30 mW/m^2$; and $K = 2.5 W/Km$) is instantaneously and homogeneously attenuated such that it transects Holdaway's (1971) andalusite-sillimanite
stability field, homogeneous extension of at least 142% is required (Fig. A8.2; $\beta \geq 2.14$ for the Richardson et al., 1969 triplepoint). As extension is not instantaneous, but progressive, initial thermal relaxation (=cooling) will coincide with the later stages of extension, and $\beta = 2.42$ is only a minimum (Fig. A8.3). The lower the strain rate during the extension, and the faster the thermal relaxation (e.g. a higher conductivity, more effective heat transfer mechanisms, or energy loss through endothermic reactions and strain), the higher $\beta$ must be if a transient geotherm is to transect the Al$_2$SiO$_5$-triple-point.

In addition, the metamorphic geotherm does not generally reflect peak T/P ratios. If the temperature at the base of the lithosphere is abnormally high prior to extension, e.g. above a mantle plume or hotspot (Crough, 1979; Houseman & England, 1986; Lister et al., in press), the extension factor required for andalusite-sillimanite facies metamorphism is lower.

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**Fig. A8.2**  *P-T diagram showing the steady-state, conductive geotherm (I) as defined in the text. If instantaneous, homogeneous lithospheric extension is to result in a geotherm that intersects the Al$_2$SiO$_5$-triple-point, the extension factor, $\beta$, has to be greater than 2.42. As extension is progressive rather than instantaneous, $\beta$ has to be greater than 2.42. The position of the Al$_2$SiO$_5$-triple-point is after Holdaway (1971). If the Richardson et al. (1969) triplepoint ("R") is used, $\beta$ is minimally 2.14.*
Taking all factors into account, it is unlikely that andalusite-sillimanite facies metamorphism would result solely from a homogeneous lithospheric extension of less than about 3, unless the pre-extensional temperature at the base of the crust is abnormally high. Such large extensions are likely to be rare in intracontinental rifts, and are considered typical of the outer parts of the continental slope on passive margins (Le Pichon et al., 1982). Heterogeneous extension, as results for example from the detachment model (Wernicke, 1985; Lister et al., 1986a, in press), may severely raise and/or compress mantle isotherms beneath relatively unstretched crust, resulting in greater heat input at lower local crustal extensions. It seems likely that such heterogeneous extension combined with magmatic heat caused the low-P facies metamorphism which occurs e.g. in the metamorphic core complexes of the western U.S..

The second constraint on the extensional model for low-P metamorphism lies in the relatively short time-scale for thermal relaxation (McKenzie, 1978). The conductive relaxation of a thermal perturbation in a system with a length-scale L and a thermal diffusivity $\kappa$, follows an exponential decay law with a time constant of $L^2/\pi^2\kappa$ (Carslaw & Jaeger, 1959). Thermal relaxation of a homogeneously thinned lithosphere with $L=150$ km and $\kappa=10^{-6}$m$^2$/s, would then have a time constant of 72 Ma. With $L=100$ km and $\kappa=10^{-6}$m$^2$/s, the time constant would only be 32 Ma. If the heat transfer rate through the lithosphere is increased (e.g. by advection), thermal relaxation would be even faster. Mantle and crustal melting, and subsequent magma ascent will, although increasing temperatures in the crust, reduce the time-scale for thermal relaxation even further. The heat anomaly will also be reduced as a result of the difference between the latent heat of melting and the latent heat of fusion, endothermic reactions and strain. The low-P facies metamorphism should therefore coincide with extension, or postdate it by less than a few tens of Ma.

Low-P facies metamorphism triggered by lithospheric extension could accompany subsequent compression, but only if compression follows virtually immediately on from a rapid, strong extension. In addition, lithospheric thickening during the compression lowers the average $T/P$ ratio, demanding correspondingly greater initial extension to produce the required crustal thermal anomaly. Given the widespread occurrence of low-P facies metamorphism in fold belts with a wide range of ages and tectonic settings, and commonly comprising thick, mature sedimentary sequences that may reflect substantial post-rift subsidence due to lithospheric cooling, it is very unlikely that it was everywhere immediately preceded by the extreme extension required. We have therefore sought an alternative thermal model that is directly linked to crustal/lithospheric compression.
Appendix 8

LOW-P FACIES METAMORPHISM INDUCED BY "CRUSTAL THICKENING/CONVECTIVE LITHOSPHERIC THINNING"

Previous modelling of metamorphism during shortening or compressive tectonism has been largely restricted to the case of instantaneous crustal thickening followed by isostatic rebound, which is controlled by the rate of erosion (England & Richardson, 1977; England & Thompson, 1984). These models produce andalusite/sillimanite (low-P) metamorphic facies only under extreme conditions (low conductivity, high basal heat flow, high radioactive heat input), and then only during post-thickening decompression (-\(\Delta P\), ±AT). Metamorphic data from many low-P fold belts, however, indicate that the low-P facies metamorphism is generally prograde and compressive (+\(\Delta P\), +AT) and is followed by essentially isobaric cooling (Phillips & Wall, 1981; Hobbs et al., 1984; Clarke et al., 1987; Reinhardt & Hamilton, in press).

We present an alternative thermal model based on the process of convective thinning of lithosphere that has been locally abnormally thickened (Houseman et al., 1981; Nataf et al., 1981; Fleitout & Froidevaux, 1982). Houseman et al. (1981) have argued that crustal shortening will initially be accompanied by proportional thickening of the underlying lithospheric mantle, giving rise to a cool, dense mantle root protruding into the asthenosphere. This root is gravitationally unstable, and their modelling showed that, for a reasonable range of parameters, convective forces will sweep the lithospheric root sideways and downwards, rapidly thinning the mantle lithosphere. They went on to argue qualitatively that the entire mantle lithosphere is likely to be replaced by hot asthenospheric material. Complete removal of the mantle lithosphere in this way is equivalent to an infinite extension of the mantle lithosphere beneath a thickened crust, and will give rise to a large thermal perturbation in the crust. The temperature increase in the crust strongly depends on the time-interval over which the newly-upwelled asthenospheric material supports convective motion. If the period of convection at the base of the crust is long (comparable to the "hot mode" of Bird & Baumgardner, 1981), then temperatures in the middle and upper crust could theoretically rise by as much as \(\approx 300\%\). Parts of the lower crust would melt, and, even if heat transfer in the crust is by conduction only, sillimanite stability would be reached in the lower and middle crustal levels (Fig. A8.3, curve B).

We have modelled, both analytically and numerically, the thermal anomaly resulting from such thinning of the mantle lithosphere during progressive, homogeneous crustal thickening. The modelling, which is detailed elsewhere (Loosveld, Chapter 4, Appendices 2-7), includes the effects of advective heat transport via magmas and aqueous fluids, as well as a range of crustal radioactive heat production capacities. Here, we present the results of only five experiments (labelled 1 to 5 in Fig. A8.4). In these
experiments, heat transfer during thickening is by conduction and, as there is generally ample evidence for fluid flow during deformation (Etheridge et al., 1983; Wood & Walther, 1986), by an upwards-directed single-pass fluid flow. After thickening, heat transfer is by conduction only. Figure A8.4 illustrates the dependence of the P-T-t trajectory on three factors: (1) the strain rate, (2) the duration of convection within the newly upwelled asthenospheric material at the base of the crust, and (3) the overlap in time of the mantle thinning and crustal thickening.

Fig. A8.3  
Schematic temperature-time diagram, showing thermal relaxation of a point on the Al₂SiO₅-triple-point depth-level (13.9km). T=242°C is the steady-state temperature at this level, as calculated from the parameters from Table A8.1. A1: instantaneous, homogeneous extension at t=0 by a factor β=2.42; A2: progressive, homogeneous extension by the same β=2.42; B: progressive crustal thickening accompanied by lithospheric thinning; C: underplating of a thick andesitic-basaltic magma at the base of the crust; D1: intrusion of a felsic magma at mid-crustal levels; D2: intrusion in the middle crust of a differentiated granitoid magma, highly enriched in heat producing elements. The bold line represents the Al₂SiO₅-triple-point temperature. The position of the break in the time scale is arbitrary. The position of the Al₂SiO₅-triple-point is after Holdaway (1971).

The first factor is elucidated by comparing the P-T-t paths of experiments 1 and 2 (strain rate \(\dot{\varepsilon}=10^{-14.3} \text{ s}^{-1}\): fast deformation) with respectively those of 3 and 4 (identical experiments but \(\dot{\varepsilon}=10^{-15.3} \text{ s}^{-1}\): slow deformation). In all four experiments upwelling of asthenospheric material to the base of the crust accompanies the early stages of thickening. In order to transect the andalusite-sillimanite field, it is necessary that \(\Delta T/\Delta P\) is initially high. The lower the strain rates, the higher the initial \(\Delta T/\Delta P\) ratios.
Selected results from 5 numerical experiments simulating homogeneous, progressive thickening of the crust and associated thinning of the mantle lithosphere. Heat transfer in the crust during thickening is by conduction and by an upwards-directed single-pass fluid flow. The effective fluid flow in experiment 1 and 2 ($\varepsilon=10^{-14.3} \, \text{s}^{-1}$: fast deformation) is $2.5 \times 10^{-11} \, \text{m} \, \text{s}^{-1}$, in experiments 3, 4 and 5 ($\varepsilon=10^{-15.3} \, \text{s}^{-1}$: slow deformation) $2.5 \times 10^{-12} \, \text{m} \, \text{s}^{-1}$.

After thickening, heat transfer is by conduction only. Experiment 1 and 3 simulated instantaneous underplating of a 75km thick and 1300°C hot slab at the base of the crust at the onset of thickening. As this slab cools by conduction during thickening, no andalusite/sillimanite conditions are reached. Experiment 2, 4 and 5 on the other hand simulated crustal thickening accompanied by instantaneous, convective thinning of the mantle lithosphere (the Houseman et al., 1981-model), with convection at the base of the crust keeping the temperature at the base of the crust high until the end of the deformational event.

In all experiments, the steady-state conductive geotherm is calculated with $K=3 \, \text{W/Km}$, $q^*=30 \, \text{mW/m}^2$, $D=10 \, \text{km}$, and $A_0=2.5 \, \mu \text{W/m}^3$. Steady-state temperatures for the thickened crust is marked by the heavy dot. Maximum temperatures reached after thickening are marked by their experiment number. The thermal diffusivity is $1.2 \times 10^{-6} \, \text{m}^2/\text{s}$. The specific density of the crust is taken as 2700kg/m$^3$. The initial crustal thickness in all experiments was 25km. Erosion and/or tectonic denudation are left out here as experiment parameters.
The second factor is elucidated by comparing the P-T-t paths of experiments 1 and 3 with respectively those of 2 and 4. Experiments 1 and 3 simulated instantaneous underplating of a 75 km thick and 1300°C hot slab at the base of the crust at the onset of thickening (a very thick slab of mafic magma). The slab, however, immediately cools by conduction during thickening, and, despite its unrealistic thickness, andalusite/sillimanite conditions are not reached. In experiments 2 and 4 on the other hand, crustal thickening is accompanied by instantaneous, convective thinning of the mantle lithosphere (the Houseman et al. 1981-model). Once emplaced, convection keeps the temperature at the base of the crust at T=1300°C until the end of thickening, resulting in crustal temperatures during the thickening event markedly higher than in experiments 1 and 3. In experiment 4 (slow deformation and constant asthenospheric temperatures at the base of the crust during crustal thickening), the andalusite/sillimanite field is transected in a prograde fashion.

If asthenospheric upwelling to the base of the crust takes place during crustal thickening, instead of at its onset, initial ΔT/ΔP ratios will be low (slightly higher than adiabatic), and the chances of transecting the andalusite and sillimanite fields on a prograde path will be reduced. Experiment 5 simulates instantaneous convective thinning of the mantle lithosphere 15 Ma. after the beginning of the crustal thickening, when the total duration of crustal thickening is 30 Ma. A larger time-proportion of the P-T-t path is characterized by kyanite-stability; andalusite is not stable at any time.

When convection at the base of the crust stops (at the end of the compressive period?), the mantle cools conductively. As the time constant for erosion (60-300 Ma.; England & Richardson, 1977) is generally greater than the thermal time constant (heat transfer by conduction only: L²/π²k; in the order of tens of Ma.), the immediately post-thickening segment of the P-T-t path will generally be characterized by isobaric to slightly decompressive cooling. The numerical experiments reveal that if lithospheric thinning is fast relative to crustal thickening, and if temperatures at the base of the crust remain high until the end of deformation, the resulting P-T-t paths will be anti-clockwise, and, depending on the original, steady-state geotherm, may transect the andalusite-sillimanite stability field in a prograde and compressive (+ΔT, +ΔP) fashion.

The immediate thermal effect of crustal shortening must in all cases be a lowering of the T/P ratios. Inherent to the "crustal thickening/convective lithospheric thinning" model is a time lag between the beginning of crustal shortening and the increase in temperature due to lithospheric thinning. Low-P facies metamorphism resulting from the model will therefore never be synchronous with early crustal shortening. The model explains why in some fold belts peak temperatures occur during the later stages of crustal thickening.
APPLICATIONS OF THE "CRUSTAL THICKENING/CONVECTIVE LITHOSPHERIC THINNING" MODEL

In order to discriminate between extensional heating and heating due to lithospheric erosion during crustal thickening in a specific fold belt, the chronology of extensional, compressional and metamorphic events must be well constrained. We are unaware of such a complete data set for any single terrane. There are, however, two Australian Proterozoic provinces (Mount Isa Inlier and Broken Hill Province) for which the data are sufficient to demonstrate that a regional low-P facies metamorphic event cannot be entirely or even largely ascribed to the preceding extensional event. Further, in both cases peak mid-crustal temperatures were apparently reached later rather than earlier in the shortening history.

In the Mount Isa Inlier, a complex Early to Middle Proterozoic history of crustal extension, sedimentation, igneous activity, crustal shortening, and metamorphism spanned the period from about 1900 Ma. to about 1500 Ma. (Blake, 1986, 1987; Page, 1983a; Page & Bell, 1986). Extensive low-P metamorphism characterized by andalusite-sillimanite facies assemblages accompanied the first and especially the second phases of a regional shortening deformation towards the end of this period. The age of the shortening event is reasonably well constrained by Rb/Sr whole rock dating between 1600 and 1550 Ma. (Page & Bell, 1986). Several tectono-stratigraphic sequences are affected by this event (Blake, 1986, 1987), and deposition of the youngest sequence was apparently triggered by crustal/lithospheric extension. Recent U/Pb zircon dating of granites and volcanics associated with the rifting brackets the extensional event between 1700 and 1670 Ma. (R.W. Page, pers. comm., 1988). Detailed stratigraphic and structural analyses have failed to demonstrate significant tectonism between the youngest extensional and the compressional events. The period between extension and peak metamorphism/shortening is therefore at least 70 Ma., and may well be in excess of 100 Ma., exceeding the time constant for thermal relaxation after lithospheric thinning.

In the Broken Hill Province (Stevens et al., 1980, in press; Phillips & Wall, 1981; Hobbs et al., 1984), recent U/Pb zircon dating and reinterpretation of Rb/Sr whole rock data indicated that the regional low-P facies metamorphism and accompanying compressional deformation postdated sedimentation by 80 to 90 Ma. (R.W. Page, pers. comm., 1988). The time between the rifting event considered to have triggered deposition of the Broken Hill sequence and metamorphism is therefore also about 100 Ma..
Anti-clockwise P-T-t paths have been petrologically reconstructed for isolated areas in both the Mount Isa Inlier and the Broken Hill/Olary Province (resp. Reinhardt & Hamilton, in press, and Hobbs et al., 1984; Clarke et al., 1987). Similar P-T-t paths are obtained in our numerical experiments which simulate crustal thickening accompanied by fast convective thinning of the entire mantle lithosphere.

CRUSTAL AND MANTLE MELTING

So far, we have neglected magmatic heat both in the extensional models and in the model which combines crustal thickening with convective thinning of the mantle lithosphere. Both lithospheric extension and asthenospheric upwelling are likely to result in mantle melting, and possible mafic intrusion into, or underplating beneath, the crust (Fig. A8.1C). This, as well as radioactive selfheating in a thickened crust, may trigger crustal melting (Fig. A8.1D). We will argue that magmatic activity, although capable of causing local low-P facies metamorphism, is not only secondary in time, but is also a second order effect which results from a primary mantle thermal anomaly whose origin must be explained.

Upwelling of hot asthenospheric material can result in the emplacement of a basaltic to andesitic underplate at the base of the crust (Furlong & Fountain, 1986) and an anomalously high heat flow in the crust. Bohlen (1987) suggested that the low-P facies metamorphism and associated anti-clockwise P-T-t paths of many granulite terranes are due to the development of a 10 to 25 km thick underplate prior to, or during, tectonic loading. Experiments 1 and 3 in Fig. A8.4, however, show that, as the lower crust buffers the heat pulse from below (even with an unrealistic 75 km thick underplate), temperatures in the middle and upper crust would not rise dramatically and, for a range of parameters, andalusite-sillimanite stability would not be reached on any level. In these experiments, underplating occurs at the onset of thickening, and heat transfer is by both advection of an upwardly migrating aqueous fluid and conduction. After thickening, the entire system cools by conduction. The initial P/T-gradient depends on the thermal relaxation rate and strain rate, but neither of the P-T-t paths transect the andalusite-sillimanite field. With lower strain rates, the anomalous basal heat flow will tend to cause a temperature-peak in the compressional segment of the P-T-t path, as thermal relaxation of a conductively cooling underplate is a relatively fast process. The thermal anomaly in the middle crust, which, if heat transfer is by conduction only, lags behind the time of underplate emplacement by some 10-20 Ma., will also be shortlived, decreasing in longevity, but increasing in intensity with depth. Crustal magmatic activity triggered by the emplacement of an underplate would only accelerate thermal readjustment.
Upwelling of asthenospheric material would result in a considerable amount of heat added to the upper mantle, and might also trigger melting of the lower crust. Crustal melting can also be caused by the higher than normal T/P ratios due to increased radioactive selfheating within a tectonically thickened crust (England & Thompson, 1984). Lux et al. (1986), based on studies in the New England Appalachians, suggested the latter mode of crustal melting resulted in regional low-P facies metamorphism.

Crustal magmatic activity, however, has essentially the same type of time and intensity constraints as extension. Once emplaced, a conductively cooling magma emits a relatively brief heat "pulse". The penetration of this heat pulse is shallow: andalusite-sillimanite conditions will only be stable within approximately 2 km of a 800°C, mid-crustal granitic slab with a thickness of 2 km (Lux et al., 1986). Also, the low-P facies conditions will only be reached during a short time-interval, as cooling and solidification of a magma is a fast process. Wells (1980) showed that, if heat transfer is by conduction alone, the readjustment of the thermal profile, after intrusion of sill-like batholiths, takes 10-20 Ma. Advection or convection of aqueous fluids in the wall rocks, and convection within the magma, would greatly accelerate thermal relaxation, so that 20 Ma. is a maximum. If crustal thickening accompanies the low-P facies metamorphism, it will, as in the case of early extension, (partially) reset the temperature rise at any level, thus further bracketing the spatial and temporal occurrence of the low-P facies metamorphic assemblages.

One other factor which could possibly affect a local heat balance is the enrichment of heat producing elements (U, Th, K), commonly observed in highly fractionated I-type granites. Loosveld (Chapter 4), however, shows that in the long-term, the relative temperature increase on any level above the granite is inversely proportional to the distance to the top of the granite. Even above a 9 km thick granite between the 18 and 27 km level, with a heat production of 5.0μW/m², sillimanite conditions will be permanently stable in a zone only =3 km thick.

Most importantly, however, both types of magmatic activity (mafic underplating and crustal melting) require a substantial thermal anomaly in the upper mantle to provide the heat for them. Both can induce prograde andalusite-sillimanite facies metamorphism, albeit of somewhat restricted temporal and spatial extent. However, we are concerned in this study with the fundamental cause of the thermal anomaly, upon which the magmatic effects can be considered to be second order.
CONCLUDING REMARKS

In many orogenic belts, lithospheric extension can be excluded as a possible explanation for low-P facies metamorphism, as both the duration of the thermal anomaly and the amount of heat released are too limited. The low-P facies metamorphic assemblages are often genetically and temporally associated with a phase of crustal thickening, and a causal link between the two should be sought. Convective thinning of the mantle lithosphere during crustal thickening results in P-T-t trajectories very similar to the ones petrologically deduced in these belts. It also explains why, after crustal thickening events, the time constants of thermal readjustment (quadratically proportional to the thickness of the conductive layer!) are much smaller (in the order of 10's of Ma.) than the time-constants expected for a thickened lithosphere (in the order of 100's of Ma.; Houseman et al., 1981; McKenzie & O'Nions, 1983). This implies that convective thinning of the mantle lithosphere may occur as a rule rather than as an exception during crustal thickening. It is anticipated that many more Hercyno-type fold belts (Zwart, 1967) will yield anti-clockwise P-T-t paths, similar to the ones recently reconstructed for a few Australian Proterozoic fold belts.

Although we have applied the modelling to convective thinning of a compressively thickening lithosphere, the calculations are equally applicable to the case of crust-mantle delamination (Bird, 1979, Bird & Baumgardner, 1981). In particular, "hot-mode" delamination (Bird & Baumgardner, 1981) accompanied by (?driving) crustal thickening has identical thermal consequences to our model, and has been used to explain the widespread low-P facies metamorphism in early Proterozoic fold belts by Etheridge et al. (1987b). We have chosen to base the thermal calculations outlined here on the "crustal thickening/convective lithospheric thinning" model because it involves a direct causal link between compressional orogeny and low-P facies metamorphism. Such a link is not demonstrated for crust-mantle delamination.