

Some causes and consequences of high-temperature, low-pressure metamorphism in the eastern Mt Lofty Ranges, South Australia

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Andalusite-, and sillimanite-bearing metapelites in the eastern Mt Lofty Ranges preserve evidence for mineral equilibration during convergent deformation at temperatures near 550–600°C and pressures of 300–500 MPa, with lateral gradients in peak temperature of about 10°C/km near the first appearance of fibrolitic sillimanite (i.e. the sillimanite-in isograd). The high-temperature portion of the metamorphic history appears to have been nearly isobaric and the inferred geometry of isograd patterns apparently established during the deformation suggests the duration of deformation to be in the range $0.3^{+0.7}_{-0.3}$ million years. These factors imply that the thermal perturbation responsible for peak metamorphism resulted primarily from localized, advective heat transfer in the lithosphere. Advection of heat in granitic magma, at least to the presently exposed crustal levels, is suggested by the intimate relationship between the attainment of peak metamorphic temperatures, convergent deformation and intrusion. The apparently short duration of deformation suggests that thermal weakening of the lithosphere during magma ascent of heat may have triggered the deformation.

Key words: advection; conduction; magmatism; metamorphism.

INTRODUCTION

Much of the interest in metamorphic rocks stems from their ability to record the effects of orogenic processes during perturbed thermal regimes. Consequently the nature of the heat sources responsible for the thermal regimes attendant with crustal metamorphism have been the subject of intense study in the last few decades (e.g. England & Richardson 1977). In high-temperature, low-pressure (high-T, low-P) metamorphic terranes the role of advective transport of magmas is of critical importance in the development of the thermal regime (Lux *et al.* 1986; Sandiford *et al.* 1991; Sandiford & Powell 1991). In this paper we assess the contribution of advective processes to the thermal evolution of a high-T, low-P terrain in the eastern Mt Lofty Ranges, South Australia (see Dymoke & Sandiford 1992). We conclude with some speculative comments on the causal relationship between advection of magmas, metamorphism and deformation in this region.

GEOLOGICAL BACKGROUND

The Mt Lofty Ranges form part of the southern Adelaide Fold Belt (Figure 1), an arcuate belt of Neoproterozoic to Lower Cambrian sedimentary rocks which were deformed and metamorphosed during the Cambro–Ordovician Delamerian Orogeny (e.g. Preiss 1987; Jenkins & Sandiford 1992). Through most of the fold belt the metamorphism has not exceeded biotite grade; however, the metasedimentary sequences of the eastern Mt Lofty Ranges have locally experienced much higher grade metamorphism, reaching migmatite grade in places and it is this region that forms the focus of this paper.

Descriptions of lithologies and comprehensive structural studies of the Mt Lofty Ranges can be found

elsewhere (Offler & Fleming 1968; Mancktelow 1990), and recent overviews of the tectonic and magmatic evolution of the region are given in Jenkins and Sandiford (1992) and Sandiford *et al.* (1992), while discussions of various aspects of the metamorphic geology are given in Sandiford *et al.* (1990), Arnold and Sandiford (1991), and Dymoke and Sandiford (1992). Consequently, we present here only a simple overview of the geology in the vicinity of the Karinya Syncline in the eastern Mt Lofty Ranges (Figure 1), concentrating on the implications for metamorphic and thermal evolution. This region is characterized by some of the highest grade rocks and steepest lateral temperature gradients in the belt. The metasediments in this region, including psammite, pelite and carbonate, form part of the Early Cambrian Normanville and Kanmantoo Groups, which record the terminal stages of sedimentation in the Adelaide Fold Belt (Jenkins & Sandiford 1992). Zircons from a tuffaceous layer in the Normanville Group have been dated by ion microprobe at 526 Ma (Cooper *et al.* 1992).

Previous investigations in the eastern Mt Lofty Ranges have shown that this area provides excellent exposures of a Buchan Style, high-T, low-P terrain with particularly well developed andalusite–staurolite-bearing assemblages (Offler & Fleming 1968; Mancktelow 1990; Dymoke & Sandiford 1992). Regional mapping of mineral assemblages (Offler & Fleming 1968; Mancktelow 1990) shows a concentric zoning of isograds around a north-northwest-trending belt of deformed, synmetamorphic, intrusives including the Rathjen Gneiss, the Palmer Granite and the Reedy Creek Granodiorite,

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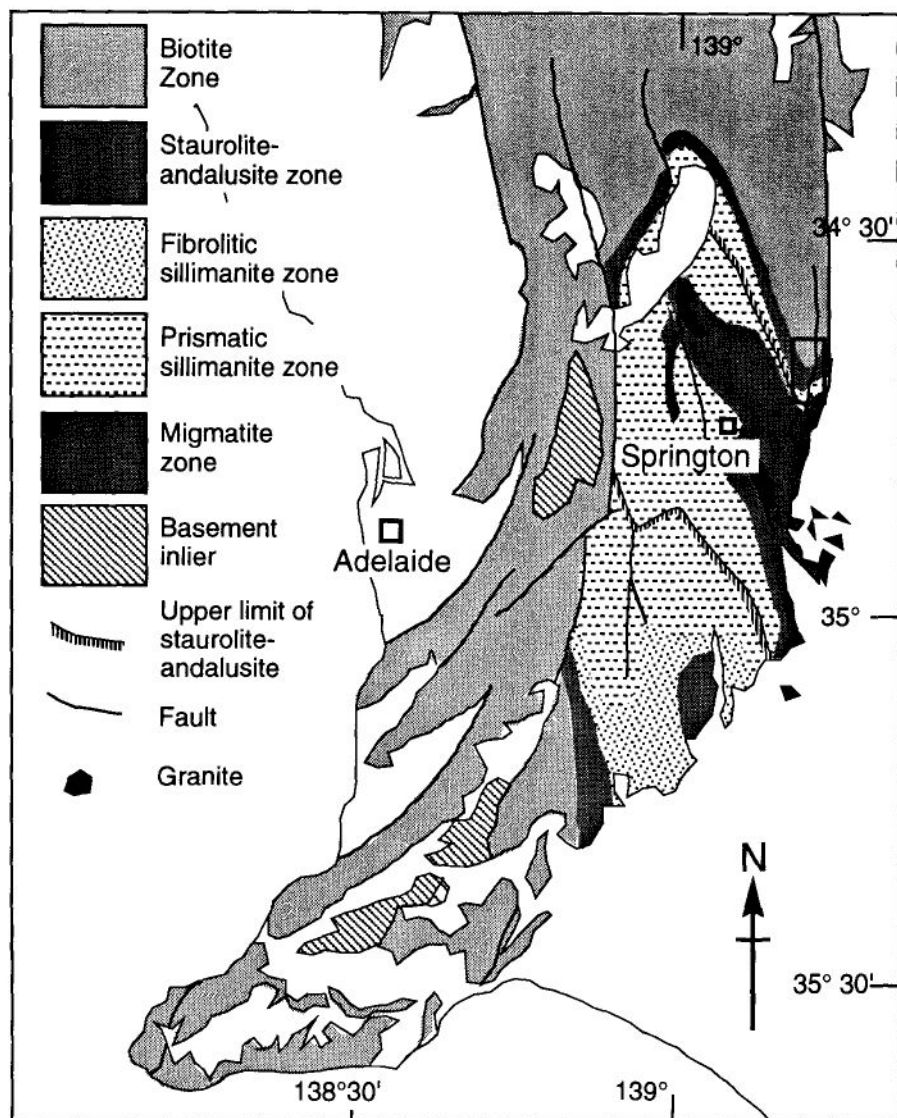


Figure 1 Geological map of the southern Adelaide Fold Belt showing the concentric metamorphic zonation in the eastern Mt Lofty Ranges (after Offler & Fleming 1968 and Mancktelow 1990). The box east of Springton encloses the area of Figure 2.

suggesting these intrusions provided at least some of the heat responsible for the metamorphism (Figure 1). The dominant structural fabric in the region of the Karinya Syncline is a steeply dipping schistosity that is best defined in micaceous schist, and is axial planar to upright to slightly reclined, tight to isoclinal, north- to north-northwest-trending folds (Figure 2). These folds and their associated axial fabrics can be traced down metamorphic grade from the highest grade zones through to biotite grade where the folds are marginally more open in style and the axial planar fabric is less well developed. In the sillimanite and migmatite zones the hinges of upright folds often contain crenulated biotite laminae, implying the existence of an earlier fabric, as does the existence of complex inclusion trail geometries in andalusite and staurolite porphyroblasts. Hence the upright folds and associated fabric are considered to be a product of the second deformation (D_2) and are here referred to as F_2 and S_2 , respectively. In detail, the regional, concentric isograd pattern can be seen to be partly folded around the longest wavelength F_2 folds, a feature which is used here to provide a constraint on the duration of folding. Evidence for the preceding deformation (D_1) is prominent

only in the highest grade zones as crenulated early fabrics in F_2 folds or as early, flat lying, migmatitic fabrics in the migmatite zone, and in the inclusion trail geometries within porphyroblasts in the andalusite–staurolite zones. At lower grades evidence for this earlier deformation is, at best, equivocal. Hence the structural complexity seems to be directly related to metamorphic grade and proximity to granitic intrusives.

The intrusive rocks in the eastern Mt Lofty Ranges form two distinct suites recognized on the basis of field relationships, geochronology and geochemistry (Foden *et al.* 1990; Sandiford *et al.* 1992; Foden *et al.* pers. comm. 1994). The older, dominantly I-type, suite was intruded prior to, and during, D_2 with granite veins either exhibiting F_2 folds or else boudinaged in the S_2 plane. Ion-microprobe ages of zircons believed to reflect igneous crystallization are in the range 515–490 Ma (Foden *et al.* pers. comm. 1994). The younger, bimodal suite of dominantly A-type intrusives are always undeformed (i.e. postdate D_2 strain) and have yielded U–Pb zircon ages of 485 Ma (Foden *et al.* pers. comm. 1994). At Reedy Creek, some 12 km south of the Karinya Syncline, U–Pb analyses of zircons from both members

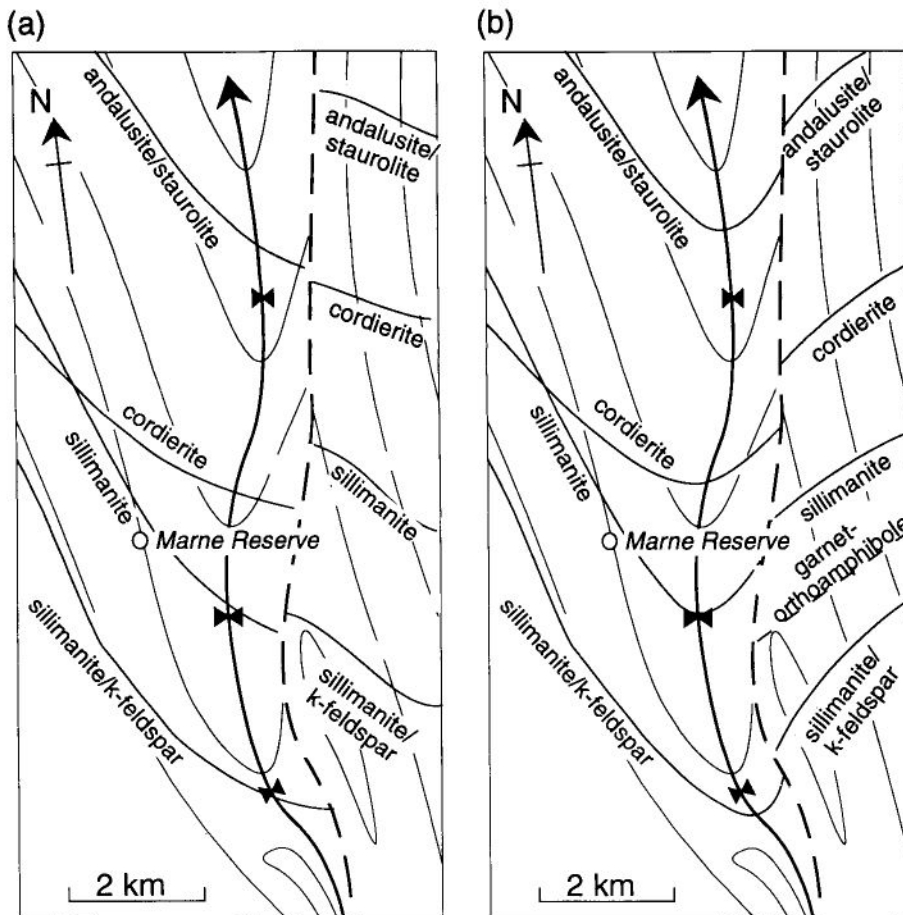


Figure 2 Inferred geometry of isograds in the Karinya Syncline according to: (a) Mills (1964) and (b) this paper. Our interpretation of the isograd geometry provides a constraint on the duration of folding in the Karinya Syncline (see text for discussion).

of the older, deformed suite (which preserves the S_2 foliation) and the younger, undeformed suite (which cross-cuts both the older intrusive as well as the S_2 foliation) are interpreted (Foden *et al.* pers. comm.) as reflecting virtually identical intrusive ages allowing, within error, for a maximum of no more than about 3 million years for the development of the S_2 foliation preserved in the older intrusives.

METAMORPHIC GEOLOGY

In this section we concentrate on the distribution and significance of mineral assemblages from the vicinity of the Karinya Syncline in the eastern Mt Lofty Ranges (Figure 2). The Karinya Syncline, with an estimated wavelength of approximately 15 km and amplitude of the order of 5 km, is a regional scale north-trending, F_2 fold which trends obliquely across the metamorphic zonation over a distance of about 40 km (Figure 2). In the south, at highest grade, the Karinya Syncline plunges gently north exposing both the stratigraphy and the regional isograd pattern in oblique profile. To the north, in the biotite zone, the fold axis flattens out eventually becoming essentially horizontal. As documented by Dymoke and Sandiford (1992), the outcrops in eastern Mt Lofty Ranges are dominated by psammitic metasediments of the Kanmantoo Group consisting of biotite, plagioclase, quartz and muscovite, and containing relatively little useful information regarding metamorphism. However, a

number of other rock types do provide critical information regarding the spatial and temporal variations in the distribution of temperatures and pressures in this terrain. These are described below.

Pelitic schist

The pelitic schists contain a variety of key index minerals including the aluminosilicates, andalusite and sillimanite, as well as K-feldspar, muscovite, chlorite, garnet and staurolite in addition to ubiquitous biotite, quartz, plagioclase and ilmenite (Dymoke & Sandiford 1992). The distribution of isograds marks the progressive increase in metamorphic grade southwards along the Karinya Syncline. The metamorphic zonation is defined, in order of increasing grade, by the following zones (Offler & Fleming 1968; Mancktelow 1990; Dymoke & Sandiford 1992): (i) biotite; (ii) andalusite + staurolite; (iii) fibrolitic sillimanite; (iv) prismatic sillimanite; and (v) migmatite.

According to Offler and Fleming (1968) the first appearance of andalusite and staurolite can be correlated with the disappearance of chlorite throughout much of the southern Adelaide Fold Belt, although we have not corroborated this observation in the vicinity of the Karinya Syncline. Near the first appearance of fibrolitic sillimanite there is considerable textural evidence for partial replacement of andalusite by sillimanite (Figure 3a) while at higher grades the aggregated habit of fibrolitic sillimanite in clusters up to 1 cm in diameter is

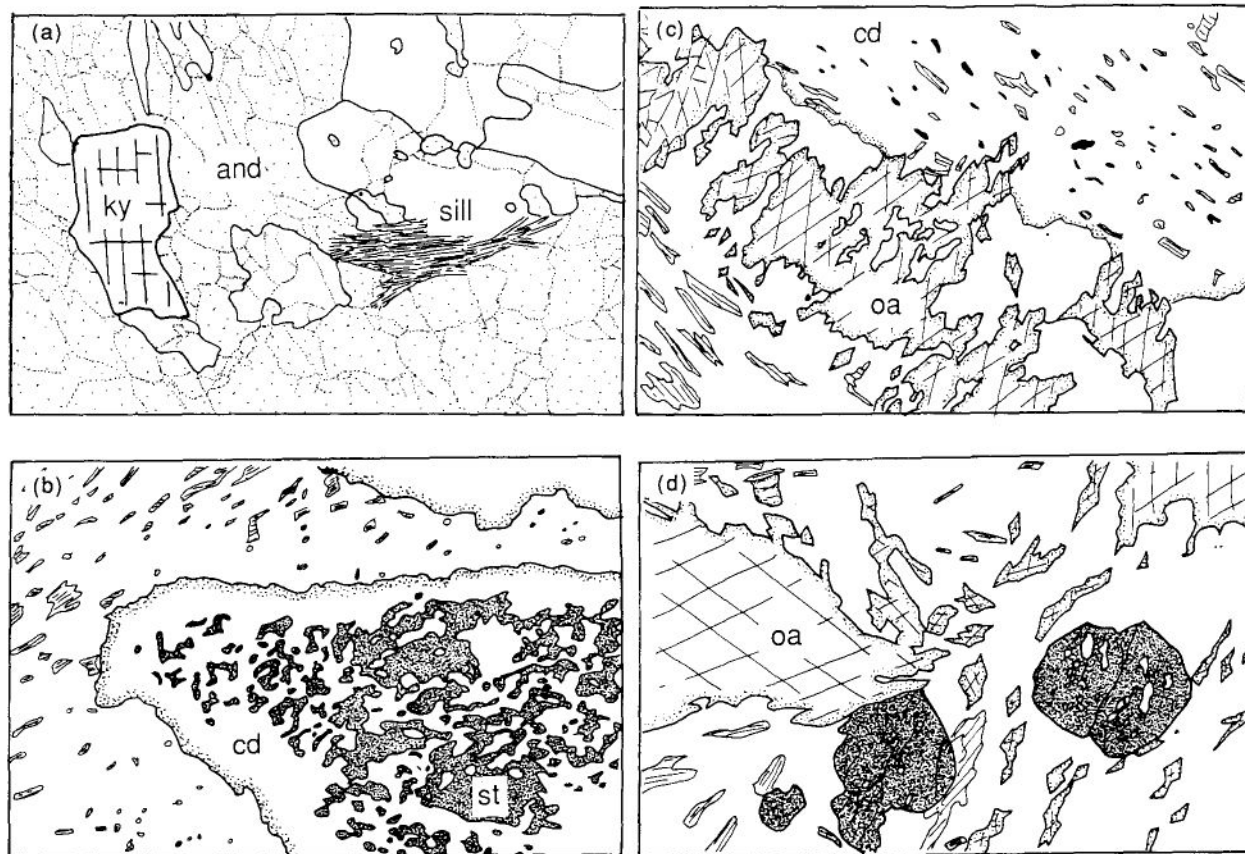


Figure 3 Tracings from photomicrographs showing (a) early kyanite replaced by andalusite and finally sillimanite in a quartz-rich segregation from near the Marne Reserve; (b) early staurolite replaced by cordierite in orthoamphibole-bearing schist; (c) late orthoamphibole growing in cordierite-bearing schist; (d) orthoamphibole–garnet schist. Field of view is approximately 3 mm wide.

suggestive of its growth being due to the breakdown of andalusite porphyroblasts. Below the prismatic sillimanite zone garnet is typically spessartine-rich particularly in association with andalusite and staurolite (Dymoke & Sandiford 1992). Apart from extensive late muscovitization there is little in the way of important retrograde mineral growth, although we have found one occurrence showing partial replacement of a garnet–sillimanite–biotite assemblage by fine-grained staurolite–chlorite–muscovite intergrowths.

Because of the general paucity of pelitic outcrops the geometry of the isograd outcrop pattern cannot be precisely defined. Mills' (1964) interpretation of the approximate position of the andalusite-in and staurolite-in isograds (mapped together), the sillimanite-in isograd and the sillimanite + k-feldspar-in isograds is shown in Figure 2a (see also Sandiford *et al.* 1990). In this interpretation the isograds cross-cut the Karinya Syncline obliquely with a prominent left-lateral displacement along a fault on the eastern limb which is essentially parallel to stratigraphy as shown in Figure 2a. Our reinterpretation of Mills' (1964) observations around the Karinya Syncline is illustrated in Figure 2b and shows that isograds are more probably folded, with a wavelength comparable to, but an amplitude about half, the folded stratigraphy. The main observational support for our interpretation of the isograd sequence is provided by the distribution of orthoamphibole-bearing assem-

blages in K_2O -deficient schists on the eastern limb of the syncline (see below), which define a northeast-trending isograd sequence (Figure 2b).

Aluminosilicate-bearing quartz segregations

Aluminosilicate-bearing quartz segregations contain all three aluminosilicate polymorphs at the Marne Reserve locality (Sandiford *et al.* 1990). Textural relations suggest early kyanite has been replaced by andalusite and finally sillimanite (Figure 3a). These segregations are up to several metres in length and are aligned parallel to the compositional layering and thus we interpret them to have formed prior to the upright D_2 folding.

Orthoamphibole-bearing rocks

Orthoamphibole-bearing rocks occur in scattered pods between the Marne River and Saunders Creek (Arnold & Sandiford 1990). In these rocks orthoamphibole occurs in association with staurolite, cordierite or garnet as well as quartz, albite and biotite. Staurolite, where present, is rimmed by cordierite (Figure 3b) in rocks which may also show late growth of orthoamphibole (Figure 3c). As described by Arnold and Sandiford (1990) these late cordierite–orthoamphibole assemblages are considered a result of the breakdown of biotite and staurolite or andalusite accompanied by metasomatic depletion of

potassium at conditions near the peak of metamorphism. However, distinct zonal isograd patterns are recognized in the distribution of assemblages in these K_2O -deficient schists, and the following progression of mineral associations with orthoamphibole is recognized from low to high grade over 5 km of outcrop (all assemblages include quartz, biotite and ilmenite): (i) staurolite + cordierite (\pm orthoamphibole); (ii) orthoamphibole + cordierite; (iii) orthoamphibole + cordierite + garnet; and (iv) orthoamphibole + garnet. An important aspect of these K_2O -deficient schists is they define the northeast-trend of isograds on the eastern limb of the Karinya Syncline (Figure 2), thus corroborating our interpretation of the distribution of pelitic isograds about the fold structure.

P-T conditions and metamorphic field gradient

Dymoke and Sandiford (1992) calculated average pressures (following the method of Powell & Holland 1988) of between 400 and 450 MPa and temperatures between 550°C and 600°C for pelitic assemblages in the andalusite-staurolite zone from the eastern Mt Lofty Ranges. Maximum temperatures in the sillimanite zone are indicated by the coexistence of sillimanite and relatively Mn-poor garnet. The $T-X_{Fe}$ relationships in

the K_2O -FeO-MgO- Al_2O_3 - SiO_2 - H_2O (KFMASH) system calculated using the dataset of Holland and Powell (1990) show that at 400 MPa a minimum temperature of about 610°C is required for this assemblage (Figure 4), although considerable sensitivity to even minor amounts of Mn is suspected (Powell, pers. comm. 1993). Conventional garnet-biotite geothermometry on garnet-sillimanite-bearing rock, using the calibration of Ferry and Spear (1978) yields temperatures of ~600–630°C.

Dymoke and Sandiford (1992) showed that the regional arrangement of isograds in the eastern Mt Lofty Ranges reflects an essentially isobaric metamorphic field gradient with maximum lateral temperature gradients (∇T_{max}) of about 10°C/km attained in the staurolite and andalusite zones near the sillimanite-in isograd.

The distribution of orthoamphibole-bearing assemblages documented above further corroborates such a metamorphic field gradient. To a first approximation, these assemblages can be understood by considering the stability of their constituent minerals in specific bulk-rock compositions in the FeO-MgO- Al_2O_3 - SiO_2 - H_2O (FMASH) system (Figure 5). Note that biotite, ilmenite and albite in these rocks are each stabilized by an additional component (K, Ti, Na, respectively) which occurs in negligible amounts in other phases.

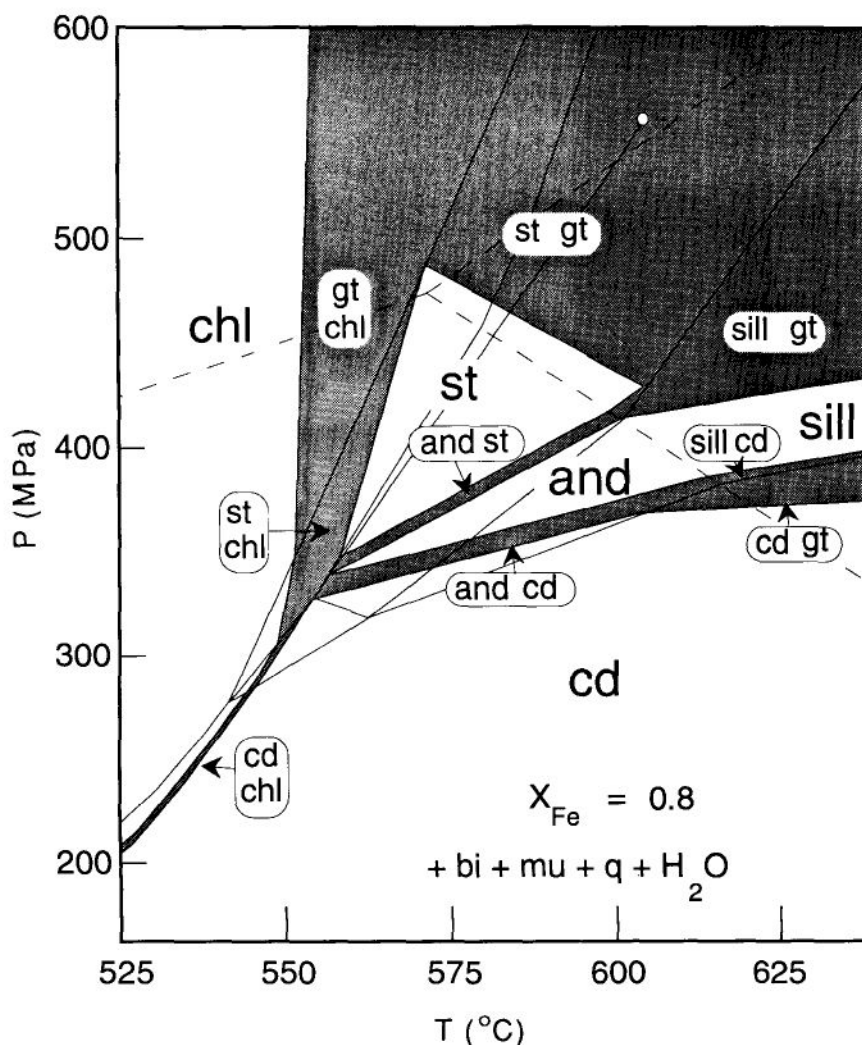


Figure 4 PT projection for the KFMASH system, appropriate to the pelitic rocks from the eastern Mt Lofty Ranges (from Dymoke & Sandiford 1992). The pseudo-section for a bulk rock composition of $X_{Fe} = 0.8$ shows the stability field of retrograde staurolite-chlorite assemblages.

and Sandiford (1992), essentially isobaric cooling (Figure 4). For an X_{Fe} appropriate to the bulk rock composition, in this case X_{Fe} of 0.8, the minimum pressure for the staurolite–chlorite assemblage is 300 MPa (Figure 4). However, the development of retrograde staurolite–chlorite assemblages but not cordierite–chlorite or cordierite–andalusite is suggestive of pressures of 400 MPa or greater during the initial retrograde P–T interval. At 300 MPa the chlorite–staurolite association is stable at about 570°C (at $a[H_2O] = 1.0$), with temperatures decreasing with increasing pressures (Dymoke & Sandiford 1992).

Distortion of isothermal surfaces by upright folding such as suggested by the partly folded isograd patterns about the Karinya Syncline must induce lateral heat transfer between adjacent fold hinges (Sleep 1979). Because the isotherms respond to the evolving structure (Figure 7), the form of isotherms developed during folding can in principle be used to constrain the duration of folding (e.g. Sleep 1979). Using various simplifying assumptions, Sleep (1979) has shown that for an upright fold that amplifies with a (vertical) velocity distribution given by:

$$V = V_z \sin(2\pi x/x_0)$$

where V_z is the maximum velocity, x is the horizontal distance, and x_0 is the wavelength of the fold. The temperature field $T_{(x,z)}$ will evolve according to:

$$T_{(x,z)} = G_0(z-z_0)(1 - \exp[-t/t_0]) \sin[2\pi x/x_0]$$

where G_0 is the initial vertical temperature gradient, and t is time. z_0 and t_0 are given by:

$$z_0 = V_z x_0^2 / (4\pi^2 \kappa)$$

$$t_0 = x_0^2 / (4\pi^2 \kappa)$$

where κ is the thermal diffusivity. Note that t_0 gives the time scale required to return to the original gradient once folding has ceased. Note that in this formalism the horizontal shortening across the fold during the amplification process is neglected; that is, the wavelength

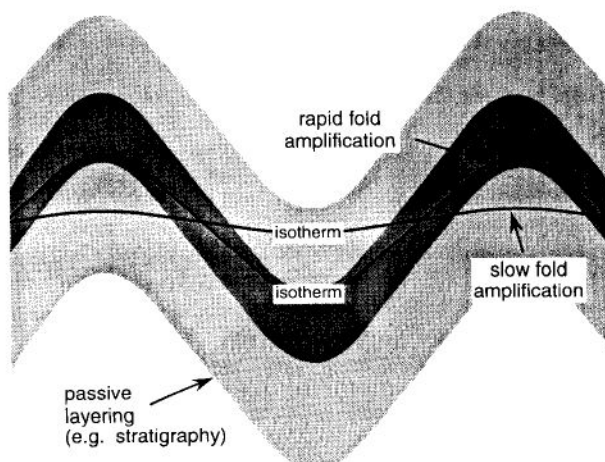


Figure 7 Schematic illustration of the evolving thermal structure around a developing fold.

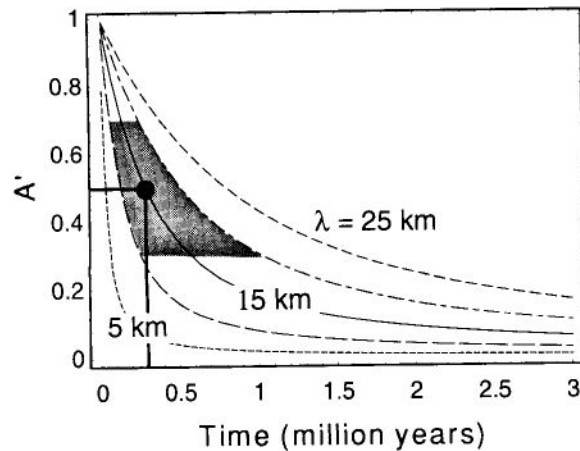


Figure 8 Application of Sleep's (1979) analysis to the geometry of isograds around the Karinya Syncline, showing the dependence of the ratio of amplitude of folded isotherms to folded stratigraphy, A' , as a function of time for fold wavelengths in the range 5–25 km (assuming a thermal diffusivity of $\kappa = 1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$). For the estimated value of $A' = 0.5 \pm 0.2$ and wavelength of $\lambda = 15 \pm 5$ km, appropriate to the Karinya Syncline, we estimate the duration of folding as $0.3 (+0.7, -0.2)$ million years.

is assumed to be constant throughout the folding episode. This assumption clearly represents a limiting case which exaggerates the thermal consequences of folding for a more general scenario in which wavelength diminishes with the amplification of the fold structure. Figure 8 shows that the ratio of the amplitude of folded isotherm to the amplitude of the folded stratigraphy, A' , depends only on the wavelength of the fold, the duration of the folding and the thermal diffusivity.

While recognizing that isograds and isotherms are not equivalent, we assume that isograd development can be equated with the general form of isothermal surfaces during the development of the Karinya Syncline. Since the mineral textures imply that the preserved isograd pattern in the Karinya Syncline was developed during folding, the geometry of isograds can therefore be used to constrain the duration of folding, with the preservation of partly folded isograd patterns implying that fold amplification occurred at a rate greater than the rate of thermal equilibration over the appropriate length-scale. For a thermal diffusivity of $10^{-6} \text{ m}^2 \text{ s}^{-1}$, as appropriate to most crustal rocks, and a wavelength of about 15 km, the estimated value of A' of 0.5 implies the duration of folding for the Karinya Syncline was 0.3 million years. An uncertainty in A' of about $\pm 20\%$ and in the estimated wavelength of the Karinya Syncline of about ± 10 km implies an uncertainty in the duration of folding of the order of $0.3^{+0.7}_{-0.3}$ million years (Figure 8). We note that independent evidence to support this result is provided by U–Pb analysis of zircons from both syn- D_2 intrusives (which preserve the S_2 foliation) and post- D_2 intrusives (which cross-cut the S_2 foliation) at Reedy Creek, as cited above, which shows that the ages of the intrusive events are virtually indistinguishable and allow a maximum time of about 3 million years for the development of the component of the S_2 foliation preserved in the syntectonic intrusives.

SOME MECHANICAL CONSEQUENCES OF METAMORPHISM

The arguments presented in the preceding sections support the notion that the thermal regimes generated during metamorphism of the eastern Mt Lofty Ranges result directly from magmatic advection of heat, and that the magmatism and metamorphism were temporally and spatially associated with a short-lived phase of convergent deformation. In terms of the mechanics of orogenic development the critical questions relate to the processes governing magmatism and whether the process of melt generation, segregation and placement significantly modified the way the orogen deformed (e.g. Sandiford *et al.* 1991; Foden *et al.* pers. comm. 1994). Since lithospheric strength is strongly temperature dependent (e.g. Brace & Kholstedt 1980), the deformation history of an orogen may also be intimately linked to the heat transport processes and will therefore depend on the rates and durations of these processes that modify the thermal regime. In particular, rapid heat transfer into the crust via magma transport may potentially trigger and localise pervasive deformation in metamorphic terrains (Sandiford *et al.* 1991). That such a relationship between deformation and thermal history is apposite for the high-T, low-P metamorphism in the eastern Mt Lofty Ranges is suggested by the geochemical signatures of magmatic rocks from the Adelaide Fold Belt which imply that the synorogenic granites are crust–mantle mixtures, and not simply the result of crustal melting (Sandiford *et al.* 1992). We suggest that the metamorphic record in the eastern Mt Lofty Ranges, in which peak temperatures were reached during an extremely short-lived, convergent phase of deformation coeval with magmatism is consistent with such a scenario. The main observations we use to support this interpretation bear on the temporal and spatial distribution of heat in the metamorphic pile as reflected in the preserved mineral assemblages, isograd patterns and textures. In particular the geometry of isograd patterns developed during folding points to extremely transient deformation. As such we regard the high-T, low-P metamorphism and deformation of the eastern Mt Lofty Ranges as a consequence of sub-crustal heat input, efficiently transported to mid–upper crustal levels with the attendant thermal weakening providing a trigger for deformation. The resolution of the nature of the mechanisms controlling sub-crustal melting remains to be elucidated.

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