The tectonic evolution of the Southern Alps, New Zealand: insights from fully thermally coupled dynamical modelling

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SUMMARY

We present here the results of numerical models of the Southern Alps in which the thermal and dynamical developments of the orogen are fully coupled to one another. This interlinkage allows the use of a wide range of thermal and physical features of the orogen as constraints on the applicability of our models. In particular, the thermal aspect of this method enables the prediction of the distribution of thermochronological ages at the surface of the model.

Perturbation of the geothermal structure due to rapid uplift and exhumation causes an intrinsic weakening and concentration of strain along the equivalent of the Alpine Fault zone in these models. This thermal weakening of the crust also produces a zone of high strain antithetic to the Alpine Fault in the upper crust that is comparable to the Main Divide Fault Zone within the Southern Alps. The introduction of orographic rainfall and erosional processes into the model leads to the development of high, asymmetric topography comparable to that of the Southern Alps. This topographic profile results from the capture of available precipitation by the windward side of the orogen and the resultant rain shadow on the leeward side. Due to the time required to accumulate sufficient topography for this rain-capture effect to become significant, the establishment of this high, asymmetric topography lags the initiation of surface uplift by several million years.

Comparison of the observed thermal history of the Southern Alps with that seen in models based on differing hypotheses of the tectonic evolution of the South Island shows that the present tectonic regime of the orogen most probably developed in a single rapid reorganization of plate motions at approximately 5 Ma with relative stability of the tectonic regime since that time. The variation in isotopic ages along the Southern Alps is consistent with that expected from variation in accumulated uplift and exhumation along the orogen arising from the obliquity of convergence of the Australian and Pacific plates.

Key words: dynamic modelling, Southern Alps, tectonic evolution, thermochronology.

1 INTRODUCTION

The rapid uplift and denudation typical of active orogenic regions exhume mid- to lower-crustal rock faster than the heat advected with it can be diffused away into the surrounding region (e.g. Clark & Jäger 1969; Allis et al. 1979; Koons 1987; Batt & Braun 1997). This leads to high temperatures at shallow depths, and a correspondingly high heat flow and near-surface geothermal gradient (e.g. Clark & Jäger 1969; Ingham 1995; Shi et al. 1996). Such thermal disturbance in turn influences the physical development of the orogen through the strong dependence of mineral strength [and corollaries such as the concept of a brittle–ductile transition in the crust, defined here as the transition from Coulomb behaviour in the upper crust (Byerlee 1978) to power-law behaviour (Kirby 1985) at depth] on temperature (Tsenn & Carter 1987; Carter & Tsenn 1987; Koons 1987; Holm et al. 1989).

These relationships are investigated here using observations from the Southern Alps of New Zealand. This prominent mountain range results from ongoing oblique thrusting of Pacific Plate crust over the Australian Plate along the Alpine Fault (Fig. 1), and distributed deformation of the Pacific Plate
up to 200 km to the southeast (Norris et al. 1990). In the approximately 5 Myr since this convergence began (Sutherland 1995, 1996; Batt et al. 1999), an estimated 50 ± 35 km of shortening has been accommodated across the plate boundary zone (Walcott 1984), with recent gravity modelling (Stern 1995) favouring the higher end of this range, around 80 km.

Despite the extent of shortening in this brief interval, and the correspondingly high uplift rates of the Southern Alps (Wellman 1979), comparable rates of erosion have prevented significant overthrusting of Pacific Plate crust onto the Australian Plate, and resulted in the exhumation of garnet-bearing amphibolite facies schists close to the Alpine Fault (Koons 1989; Norris et al. 1990). The rapidity of this regional exhumation has had a profound influence on the thermal regime of the Southern Alps, with many geophysical indicators pointing to anomalously high near-surface temperatures beneath the region (e.g. Craw 1988; Holm et al. 1989; Craw et al. 1994; Ingham 1995; Allis & Shi 1995; Shi et al. 1996).

This paper presents the results of selected finite element models investigating the tectonic evolution of the Southern Alps, using recently collected geochronological data (Tippett & Kamp 1993; Batt 1997; Batt et al. 1999) as a principal constraint. This work represents an advance over previous modelling of the development of the Southern Alps (e.g. Allis et al. 1979; Koons 1987; Shi et al. 1996; Beaumont et al. 1996) due to the fully integrated modelling of thermal and physical evolutions given here. Coupling of this sort, and in particular the incorporation of the diffusion of heat (cf. Beaumont et al. 1996) allows the investigation of the behaviour of material relative to the evolving thermal structure of the orogen. This enables us to use thermochronology as a direct indication of the state of the evolving orogen through time. Such a constraint offers important insight into the development of the Southern Alps, and also leads to the production of models more consistent with the observed properties of the orogen.

We initially provide a first-order estimate of the extent to which the thermal perturbation observed in the Southern Alps has influenced the physical development of the orogen. The isotopic age predictions of the model are then compared with the available thermochronological data from across the orogen to test the applicability of different hypotheses of the evolution and present dynamics of the Southern Alps.

2 MODELLING

2.1 Kinematic model

In the following numerical experiments, deformation of the crust is driven by a velocity discontinuity imposed at the base of the model (Fig. 2), representing detachment and subduction of the mantle underlying one of the two lithospheric plates involved (after the method of Willett et al. 1993). As this mantle subducts, the overlying lighter crust remains buoyant, and is in turn forced to accommodate the plate convergence by deforming (Fig. 2). The deforming crust behaves as a doubly vergent accretionary wedge, the geometry of which is determined by the internal strength of the crust and the strength of the detachment between the crust and mantle (Dahlen & Barr 1989; Barr & Dahlen 1989). The side of the model on which mantle subduction is referred to as the pro-side of the orogen, and the side on which the ‘stable’ mantle resists subduction is defined as the retro-side (after Willett et al. 1993).

Following the convention of Beaumont et al. (1996) and Shi et al. (1996), the mantle in all models discussed in this study
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Figure 2. (a) Simplified description of the tectonic setting assumed in our numerical models. Subduction of the subcontinental mantle is accommodated by thrusting in the overlying crust. (b) Schematic description of the numerical model and boundary conditions used to represent the tectonic setting described in (a).

converges from the right (the pro-side), while the left-hand (retro) side of the model is assumed to remain stationary. The boundary between these ‘plates’ is assumed to be fixed with respect to the stationary retro-side of the model. This gives the same sense of motion with respect to the Southern Alps as Pacific Plate convergence relative to a fixed Australian Plate when viewed from the south.

2.2 Numerical method

The crust is modelled in two dimensions within the outlined kinematic bounds, deforming as a non-linear Maxwell visco-elastic body until stress levels reach a critical value and brittle failure takes place. The viscosity of the material is stress- and thermally activated, allowing temperature to feed back into the model’s mechanical properties. In this respect, rock viscosity follows a ‘classical’ Arhenius relationship derived from laboratory analysis of rock samples

\[ \dot{\varepsilon} = A \sigma^n e^{Q/RT}, \]  

where \( \sigma \) is differential stress, \( \dot{\varepsilon} \) is strain rate, \( A \), \( n \), and \( Q \) are material flow parameters, \( T \) is the absolute temperature and \( R \) is the gas constant (Goetze & Evans 1979; Carter & Tsenn 1987; Tsenn & Carter 1987).

Two rheologies are considered in this work: quartz-dominated material (R1) based on the quartzite rheology of Paterson & Luan (1990) for low- to moderate-grade geological conditions (with \( A = 6.5 \times 10^{-8} \text{ MPa}^{-n} \text{ s}^{-1}, n = 3.1 \), and \( Q = 135 \text{ kJ mol}^{-1} \)), and feldspar-dominated material (R2) based on the Adirondack granulite rheology of Wilks & Carter (1990) (\( A = 8 \times 10^{-3} \text{ MPa}^{-n} \text{ s}^{-1}, n = 3.1 \), and \( Q = 243 \text{ kJ mol}^{-1} \)). Due to an absence of strain-weakening behaviour in the model, these figures should be regarded as giving absolute maxima for the strength of the modelled crust.

Brittle deformation is represented by an associative plastic flow law derived from Griffith’s failure criterion (Griffith 1921; Murrel 1963). This failure criterion can be expressed in terms of the stress tensor,

\[ J_{2D} + 12T_0 p = 0, \]  

where \( J_{2D} \) is the second invariant of the deviatoric part of the stress tensor, \( T_0 \) is the tensile strength of the crustal material and \( p \) is the pressure (Murrel 1963). In the model used here, this pressure term incorporates the lithostatic pressure (arising from the weight of the overburden) and the dynamical pressure (resulting from deformation driven by the imposed boundary conditions). Braun (1994) and Braun & Beaumont (1995) have previously shown that the use of the Griffith failure criterion to model the brittle deformation of crustal rocks produces results that agree qualitatively with both observations of fault orientation in nature and the results of sandbox analogue experiments, and also quantitatively with direct measurement of dilatancy during high-pressure rock deformation experiments.

The deformed crustal layer is assumed to rest on a relatively strong mantle lithosphere of finite flexural strength. This is incorporated in the model by linking the deformation at the base of the crustal layer to a thin elastic plate which supports loads created by variations in the thickness of the crustal layer (Table 1).
Table 1. Values of the physical and thermal parameters that remain constant throughout the modelling experiments.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crustal thickness, L</td>
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<tr>
<td>Young's modulus, E</td>
<td>$10^{10}$ Pa</td>
</tr>
<tr>
<td>Poisson's ratio, ν</td>
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<tr>
<td>Tensile strength, $T_0$</td>
<td>$10^{10}$ Pa</td>
</tr>
<tr>
<td>Time step length, $\Delta t$</td>
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<td>Thermal diffusivity, $\kappa$</td>
<td>25 km$^2$ Ma</td>
</tr>
<tr>
<td>Elastic thickness of lithosphere, $z_e$</td>
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<tr>
<td>Mantle density, $\rho_m$</td>
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</tr>
<tr>
<td>Crustal density, $\rho_c$</td>
<td>2900 kg m$^{-3}$</td>
</tr>
</tbody>
</table>

Geometrical effects arising from finite crustal deformation are also incorporated. The model calculations use the midpoint strain in place of the infinitesimal strain and the Green–Naghdi rate of stress change in place of the time derivative of stress (after Hughes & Wingett 1980).

The overall deformation of the model is constrained by the requirement of mechanical equilibrium, such that

$$\frac{\partial \sigma_{ij}}{\partial x_j} = -\rho g \delta_{ij}, \quad (3)$$

where $\rho$ is crustal density, $g$ is gravitational acceleration, and the subscript 2 refers to the vertical dimension.

For those experiments where topography is allowed to develop, erosion and sedimentation rates along the upper surface of the model are computed using a 1-D model of surface processes similar to that developed by Kooi & Beaumont (1994).

Heat transport in such tectonically active regions is governed by the equation:

$$\frac{\partial \mathbf{T}}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{T} = \kappa \nabla^2 \mathbf{T} + \mathbf{Q}, \quad (4)$$

where $\mathbf{u}$ is the tectonic velocity measured in a system of reference attached to the retro- (or fixed) side of the orogen, $T$ is temperature, $\kappa$ is the thermal diffusivity of the crust (Table 1) and $\mathbf{Q}$ is any internal heat source. Due to the relative lithological homogeneity of the crust making up the Southern Alps (e.g. Reynolds & Cowan 1993; Smith et al. 1995), radioactive heat generation is assumed to be uniform for the crust in the model (Table 1), with no radiogenic heat source for the mantle. The possible contribution of frictional heating to the geothermal regime of the Southern Alps has long been an area of debate (see Shi et al. 1996 for a discussion), and this factor is not included as a heat source in the models presented here.

The equations of mechanical equilibrium (eq. 3) and heat transfer by advection and conduction (eq. 4) are solved by the finite element method (Bathe 1982) at 20000 yr time intervals using linear triangular elements. To permit large deformation, the ‘Dynamic Lagrangian Remeshing’ (DLR) method is used (Braun & Sambridge 1994). The DLR method was further improved for application to this problem through allowing for dynamical mesh refinement, as described in Batt & Braun (1997).

A constant-temperature boundary condition is assumed at the base of the model, neglecting the cooling effects of the lower lithospheric subduction inferred to be taking place beneath the orogen (Fig. 2). This simplification has been shown to be justified in the case of young orogens such as the Southern Alps (<10 Myr duration) that lack a prolonged history of subduction. The cooling effects of subducted material have insufficient time to be felt in the upper crust on such timescales (Shi et al. 1996; Batt & Braun 1997).

Table 2. Convergent and strike-slip components of relative Australian–Pacific Plate velocity along the Alpine Fault, calculated assuming that the Australian Plate is stationary and motion of the Pacific Plate is described by motion about the pole of rotation discussed in the text on the surface of a sphere of radius 6371 km. The Alpine Fault is assumed to have a constant orientation of 055° (after Norris et al. 1990). Note that these calculations do not allow for strain partitioning, assuming 100 per cent accommodation of the relative plate velocity on the Alpine Fault, and should therefore only be taken as a guide.

<table>
<thead>
<tr>
<th>Latitude (°S)</th>
<th>Longitude (°E)</th>
<th>PAC–AUS relative motion magnitude (mm a$^{-1}$)</th>
<th>PAC–AUS relative motion azimuth (°E of N)</th>
<th>Deviation of Alpine Fault from plate motion azimuth (degrees)</th>
<th>Convergent component of PAC–AUS velocity vector (mm a$^{-1}$)</th>
<th>Strike-slip component of PAC–AUS velocity vector (mm a$^{-1}$)</th>
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3 OBSERVATIONAL CONSTRAINTS ON MODELS

3.1 Thermochronology

At any given locality within the Southern Alps, significant variation exists between the apparent ages of different thermochronometers, with systems of higher closure temperature [the temperature below which formerly open-system diffusion of the relevant radiogenic product out of a sample ceases, so that the daughter product in question begins to ‘age’ for the chronometer concerned (Dodson 1973)] consistently giving older ages than mineral systems from the same rocks with lower closure temperatures (Tippett & Kamp 1993; Batt 1997; Batt et al. 1999). Close to the Alpine Fault, apatite and zircon fission track ages, and muscovite, biotite, and in some cases hornblende, K–Ar ages are low (generally < 5 Ma) and relatively consistent. Beyond some critical distance from the Alpine Fault, ages then increase dramatically with further distance to the southeast, reaching early Cenozoic (for apatite fission track ages) or Mesozoic (for other chronometers) levels over a distance of several kilometres (Fig. 3). The distance from the Alpine Fault to this transition from low, consistent ages to rapidly increasing ages is related to closure temperature, with muscovite K–Ar ages displaying the marked increase in age closer to the Alpine Fault than biotite K–Ar ages, which in turn exhibit the transition closer to the Alpine Fault than zircon and then apatite fission track ages. Beyond this transitional zone to the southeast, ages remain at these higher early Cenozoic to Mesozoic levels, but exhibit significant variation (Fig. 3).

Plotting all the isotopic ages which fall within the zone of young material close to the Alpine Fault (as discussed above) against distance along the Southern Alps illuminates a regular variation in age along the orogen (Fig. 4). Between Fox Glacier and the Whataroa River, the youngest K–Ar ages observed are approximately 1 Ma for biotite and 1.5–2 Ma for muscovite (Batt 1997). Adams (1981) has also reported a scattering of older ages (up to 5 Ma) from Fox River and Dry Creek in this area (Fig. 4). North of the Whataroa River, ages increase gradually with distance northeast along the orogen, with the youngest K–Ar biotite ages observed increasing from 1 Ma at Whataroa (Batt 1997), to 2.1 Ma at Mikonui River (Hawkes 1981) and 3.3 Ma at Hokitika River (Batt 1997).

South of the Fox River Valley, K–Ar ages increase approximately linearly towards Haast. Ages of approximately 4 Ma for biotite and 5 Ma for muscovite along the Copland Valley increase to 6 Ma for biotite and 7 Ma for muscovite at Mahitahi River (Batt 1997), with a well-constrained 40Ar/39Ar muscovite age of 9 Ma from Paringa River (Fig. 4) further to the south (Chamberlain et al. 1995). This trend is disrupted by the effects of excess argon on apparent ages from the Mataketake Range (Batt et al. 1999), Mt Kinnaird (Batt 1997) and Haast Pass (Adams & Gabites 1985).

Reset zircon fission track ages (where ‘reset’ refers to a mineral chronometer having lost all connection to its crystallization and previous thermal history, so that it records only cooling related to the present phase of tectonism in an area) close to the Alpine Fault exhibit a similar variation pattern, offset somewhat to the south (Fig. 4), decreasing linearly northwards in the far south of the Southern Alps, remaining relatively consistent across the central region of the orogen, and increasing slightly again in the far north.

Limited geochronological constraints are also available for the region to the northwest of the Alpine Fault. In South Westland, Kamp et al. (1992) report that apatite fission track ages of less than 5 Ma extend up to 10 km northwest of the Alpine Fault, with ages then increasing rapidly with further distance from the plate boundary. In contrast, zircon displays consistent Mesozoic fission track ages throughout this region (Kamp et al. 1992). This indicates that denudation northwest of the Alpine Fault in South Westland during the ongoing orogenic phase has exhumed material from below the fission track closure temperature of apatite (deeper than 4–5 km; Tippett & Kamp 1993), but has been insufficient to exhume material from the higher-temperature annealing zone of fission tracks in zircon (approximately 10 km depth; Tippett & Kamp 1993).

In North Westland, material has been exhumed from below the fission track closure temperatures of both apatite and zircon during this recent orogeny (Kamp et al. 1992), but not from depths and temperatures great enough to reset the K–Ar age of biotite in the same region (Rattenbury 1987).

### 3.2 Physical constraints

The physical nature of the Southern Alps has largely been summarized in Beaumont et al. (1996). Additional aspects are detailed here, however, particularly as regards to the timing of the development and thermal character of the orogen.

A segmented series of thrust and transfer faults known collectively as the Main Divide Fault Zone (Cox & Findlay 1995) occurs southeast of the Alpine Fault in the deformed Pacific Plate. The regional trend of this fault system runs parallel to and 10–20 km southeast of the Alpine Fault, with thrust segments of the Main Divide Fault Zone dipping anti-parallelly towards the Alpine Fault at 40–60°NW (Cox & Findlay 1995). Due to this relationship to the regionally dominant Alpine fault structure, the rocks of the Main Divide Fault Zone are being continually uplifted and carried westwards over time (and thus being removed from the site of active deformation, preventing the accumulation of significant finite slip on individual fault surfaces), while new faults form in the position previously occupied by the fault zone (Cox & Findlay 1995). The present exposures of this feature are thus only the most recent expression of this structural feature, with some schist fabrics west of the Main Divide Fault Zone possibly representing deeper equivalents of old Main Divide faults now uplifted and exposed by erosion (Cox & Findlay 1995).

The climatic conditions experienced across the Southern Alps are strongly asymmetric. The Southern Alps present a barrier to the predominant westerly winds of the region, resulting in orographic rainfall in the west of the orogen and a corresponding rain shadow effect and dry conditions to the east. This process gives exceptionally high rainfall of up to 12 m a⁻¹ to the west of and in the high peaks of the Main Divide (Griffiths & McSaveney 1983).

West of the Main Divide, mean surface elevation drops off steeply to the Alpine Fault, and surface elevation is unrelated to lithological variation or to the total rock uplift experienced in a given area (Tippett & Kamp 1995). In contrast, east of the Main Divide the regional slope is more gentle and the orogen broader, with elevation proportional to the total amount of rock uplift experienced in a given location (Tippett & Kamp 1995). From a substitution of time for space in the interpretation of fission track age trends across the Southern Alps, Tippett & Kamp (1995) suggest that significant surface uplift (and the resultant development of this major topography) lagged behind the initiation of rock uplift east of the Main Divide by approximately 2 Myr.

Sedimentary basins are present to both the west and the east of the Southern Alps. To the west, in the Westland and Taranaki basins, a prograding sequence of clastic sediments some 50–100 km wide, 700 km long and 3–4 km thick developed in the Pliocene and Quaternary (Sutherland 1996). Recent uplift associated with the ongoing orogeny has deformed sediments at the eastern margin of the Westland basin into a westward-dipping monocline (Kamp et al. 1992; Nathan et al. 1986). This active deformation and accompanying erosion has removed much of the terrestrial sedimentary record of the development of the Southern Alps from the West Coast region (Kamp et al. 1992). An exception to this is the Rappahannock Group in the fault-bounded Maruia Basin immediately west of the Alpine Fault in northern Westland (Fig. 1). This group is wholly terrestrial in origin, and composed dominantly of coarse conglomerates, with sedimentary provenance thus relatively easy to assess (Cutten 1979). The appearance of clasts of quartzofeldspathic schist in these conglomerates marks the base of the uppermost unit in the group, the Devils Knob Formation (Cutten 1979).

The age of the Rappahannock Group is uncertain, with no datable material yet recovered from the coarse sediments of the unit (Cutten 1979). On the basis of lithological correlation with the Old Man Group in the nearby Grey and Inangahua Valleys, Cutten (1979) assigns a mid-Pliocene age of 3.1–3.45 Ma to the base of the Devils Knob Formation. The appearance of schist clasts at the base of this formation, however, does not mark the initiation of the uplift of the Southern Alps, but rather the arrival of the Maruia Basin adjacent to the Alpine Schist due to progressive right-lateral strike-slip movement along the Alpine Fault (Cutten 1979). The coarse, lithic-fragment-rich conglomeratic units of the lower Rappahannock Group suggest that appreciable youthful relief existed east of the Alpine Fault significantly earlier than this.

In the offshore sedimentary record, major increases in sedimentation rate observed in the Waiho-1 drillhole at 5 Ma are believed to reflect the beginning of widespread uplift of the Southern Alps at this time (Sutherland 1996), while the nature of schist clasts observed in this drillhole record indicates that amphibolite facies Alpine Schist was exposed by approximately 4 Ma (Sutherland 1996).

To the east of the Southern Alps, the Canterbury Plains are covered by a veneer of greywacke gravels up to several hundred metres thick, and offshore currents have carried sediments over great distances to the Bounty Fan and to the Hikurangi and Kermadec trenches (Field et al. 1989; Carter & McCave 1994).
Heat-flow determinations for the South Island of New Zealand are summarized by Shi et al. (1996). Although the offshore heat flow is well established from measurements in oil exploration drillholes, only two measurements have been made in the Southern Alps themselves. In a drillhole in the Franz Josef area 4 km southeast of the Alpine Fault (Fig. 1), the calculated heat flow of 190 ± 50 mW m⁻² (Shi et al. 1996) was approximately three times the average heat flow of 60 ± 15 mW m⁻² observed in the Taranaki Basin to the northwest and the Great South Basin to the southeast (Funnel et al. 1995). In the Haast area, a drillhole 6 km from the Alpine Fault (Fig. 1) gave a calculated heat flow significantly lower than that obtained at Franz Josef, 90 ± 25 mW m⁻² (Shi et al. 1996).

The numerous hot springs occurring in the central and northern regions of the Southern Alps have also been suggested to reflect elevated heat flow associated with high uplift and erosion rates (Allis et al. 1979). In their re-evaluation however, Shi et al. (1996) suggest these hot springs are not a useful indicator of thermal structure, and are more closely associated with anomalously high permeability than high near-surface temperatures.

The pressure–temperature history of several areas of the Southern Alps has been calculated from the entrapment conditions of fluid inclusions found within exposed material (e.g. Craw 1988; Holm et al. 1989; Craw et al. 1994). In the work of Craw et al. (1994) three different generations of fluid inclusions from Craig Peak near Franz Josef yield distinct trapping conditions indicating that the region underwent near-isothermal uplift from mid-crustal to shallow depths, and then cooled extremely rapidly as it approached the surface (Craw et al. 1994). Such P–T paths suggest the presence of hot material at shallow depth beneath the Southern Alps, and a correspondingly high near-surface geothermal gradient. Similar results were also reported from the Waiho Valley by Holm et al. (1989).

This style of thermal structure is also supported by the results of a magnetotelluric study across the central Southern Alps (Ingham 1995) (Fig. 1). A region of enhanced electrical conductivity exists at a depth of approximately 10 km beneath the orogen in this area. Ingham (1995) interpreted this to reflect the presence of elevated temperatures (approximately 400 °C) in the upper crust beneath the Southern Alps (Ingham 1995).

4 MODEL RESULTS

This section presents the results of a series of finite element models designed to investigate the nature and history of the Southern Alps. As shown in Fig. 5, increasing levels of complexity are progressively incorporated into these models to account for observations of the orogen's character (as discussed in the previous section).

Each model result that appears to fit with a particular aspect of the character of the Southern Alps is analysed in terms of the physical processes that lead to this agreement. This allows

Figure 5. Flowchart illustrating the design of the modelling experiments.
the gradual development of an understanding of the dominant processes responsible for the present-day state of the Southern Alpine orogen.

4.1 Model 1: basic behaviour

Total denudation (where any material uplifted above the initial model surface is completely removed from the system) is used here as the basis of model 1 (M1). This removes the influence of factors such as topography, variable erosion and sedimentation, effectively isolating the fundamental mechanical behaviour of the model (Fig. 5).

M1 has a 30-km-thick crust of a feldspar-dominated rheology (R2), with the temperature at the base of the crust assumed to be fixed at 600 °C. Under these geothermal conditions, the ductile strength of R2 material falls below its brittle strength, resulting in the initiation of viscous (i.e. ductile) deformation towards the base of the modelled crustal layer (Fig. 6). Convergence occurs at a constant rate of 10 km Ma\(^{-1}\) over an interval of 10 Ma.

Although our choice of basal temperature is notably hotter than the 350-460 °C adopted by Beaumont et al. (1996) in their models of the Southern Alps, Beaumont et al. (1996) assigned temperature purely to produce a desired level of strength at the base of the model crust. Their choice was not motivated or supported by direct (or indirect) observation (Beaumont et al. 1996, p. 3341). In contrast, our analysis of thermochronological and other constraints on the thermal evolution of the Southern Alps indicates that material presently exposed at the surface of some parts of the orogen has been exhumed from depths where temperature is significantly higher than 500 °C (the closure temperature of argon in hornblende),

![Diagram of model 1](https://example.com/model1_diagram.png)

**Figure 6.** Results of model 1. (a) and (b) Greyscale contour plots of the logarithm of the second invariant of the deviatoric part of the strain rate tensor after 6 and 10 Myr of tectonic deformation, respectively. The arrows indicate the instantaneous particle velocities at selected grid points, while the thin dashed line marks the approximate location of the brittle-ductile transition in the model. (c) and (d) Greyscale contour plots of temperature after 6 and 10 Myr of tectonic deformation. Thin black lines are stratigraphic horizons that were horizontal prior to the onset of tectonic activity.

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necessitating the higher basal temperature adopted here. This
temperature of 600 °C at 30 km is also consistent with the
independent estimates of Funnel et al. (1995), who proposed
a temperature of 800 °C at 40 km depth from a regional heat-
flow study of the Tarariki Basin to the northwest and the
Great South Basin to the southeast of the Southern Alps.

Despite this thermal difference, the basic behaviour observed
in M1 follows that described by Beaumont et al. (1996) in many
respects. During the initial stages of deformation, two conjugate
shear zones develop, rooted at the velocity discontinuity at the
base of the model and dipping at approximately 45° in opposite
directions (Fig. 6). These shear zones are referred to with the
prefixes pro- and retro-, depending on the side of the orogen
on which they occur (after Willett et al. 1993).

Over time, the system evolves towards an asymmetric
distribution of deformation (Fig.6a) as the retro-shear zone
accumulates strain and develops into a major crustal-scale
thrust structure. Although the pro-shear zone also remains the
locus of concentrated instantaneous deformation, material is
rapidly advected through this feature in the frame of reference
of the stationary retro-side of the model, and thus the pro-shear
zone does not accumulate significant finite deformation.

Exhumation in the model is concentrated between these
major shear zones, reaching a maximum at the retro-shear zone
and, once steady state is achieved, decreasing approximately
linearly towards the pro-shear zone (Fig 6d). Significant
exhumation is also observed retrowards of the main deforming
wedge, again decreasing with distance from the retro-shear
zone. This behaviour reflects shear or ‘fault drag’ along the
 retro-shear zone, and could account for the minor uplift and
exhumation observed on the West Coast of the South Island
 to the northwest of the Alpine Fault (Kamp et al. 1992).

Where M1 departs from the results of Beaumont et al.
(1996) is in the consequences of the progressive evolution of
the temperature field of the model (Figs 6c and d). At the
rapid rates of uplift and exhumation experienced here, material is
exhumed much faster than the heat advection with it can be
conducted away into the surrounding crust (e.g. Clark & Jäger
1969; Allis et al. 1979; Koons 1987; Batt & Braun 1997). Given
the temperature-dependent viscous deformation experienced
at the base of M1, this thermal evolution significantly alters
the physical properties of material in the deforming region—
effectively raising the transition from elastic to plastic defor-
 mation (the ‘brittle–ductile transition’) closer to the surface, as
discussed by Koons (1987). The greatest increase in temper-
ature at a given depth in the crust, and hence the greatest
decrease in strength, is experienced along the retro-shear zone,
which is bringing hot material from the base of the crust to the
surface (Fig. 6). This relative weakening of the retro-shear
zone increases the concentration of strain accommodated on
that structure relative to the pro-shear (Fig.6a,b), increasing the
asymmetry of the model.

Despite this thermal weakening and focus of strain, the retro-
shear zone remains a relatively broad feature in this model,
reaching a minimum of several kilometres thick and showing
no sign of narrowing towards the surface. This behaviour shows a marked lack of agreement with the observed nature of the
Alpine Fault (the inferred retro-shear structure of the
Southern Alps), which is effectively a discrete feature at the
 surface, and widens to a shear zone estimated to be no more
than approximately 1 km wide at depth Sibson et al. (1979).
This lack of fit probably arises in large part because the model
does not incorporate strain-softening effects, and therefore
cannot produce narrow zones of deformation comparable to
those seen in nature. In effect, the highly strained material of
the retro-shear zone in this model has the same mechanical
strength as undeformed crustal material at the same temper-
ature, thus neglecting the influence of processes that can act
to reduce the strength of deformed rocks in nature.

Where the basal temperature of the model is reduced to
400 °C, purely frictional deformation ensues throughout the
crust, and the model results become much more symmetrical
in character, and closer to those of Beaumont et al. (1996). A
further consequence of the rheological evolution of M1 is the
development of a secondary pro-shear zone within the
deforming region (Fig.6a). As the brittle–ductile transition
migrates upwards, the pro-shear zone at first broadens, and
then at around 4 Myr into the model’s evolution begins to
be resolved into two discrete structures. One of these, the
main pro-shear zone, remains rooted at the basal velocity
discontinuity, while the other is rooted at the intersection of
the retro-shear zone with the brittle–ductile transition, and
continues to migrate upwards until this rheological boundary
attains an equilibrium position in the deforming region at
around 6 Myr into the model’s evolution.

This behaviour could account for the origin of the Main
Divide Fault Zone (MDFZ) within the Southern Alps (Cox &
Findlay 1995). This structure dips antithetically to the Alpine
Fault and is inferred to intersect it at 10–12.5 km depth. Cox
& Findlay (1995) interpreted the MDFZ as a backthrust
(or pro-shear in the terminology adopted here) structure
reflecting upper-crustal detachment at this 10–12.5 km level,
but no other evidence of such a shallow-level detachment is
seen in the geophysical signature of the orogen. As shown
here, a secondary pro-shear structure conforming to this
description can result purely from thermal perturbation of the
crust and related uplift of the brittle–ductile transition.

The spatial and temporal evolution of the temperature field
within the model also allows the evaluation of the pressure–
temperature–time (P–T–t) path for any desired point. For
material exposed at the surface of the model orogen, such
histories are characterized by an extended period of near-
isothermal cooling as material is brought up from its initial
depth through the perturbed thermal structure of the orogen,
followed by a brief episode of extremely rapid cooling as it
near the surface and encounters the high geothermal gradients
of the surface boundary layer (Fig. 7a). This history coincides
well with that obtained from fluid inclusions for the Franz
Josef area in the work of Craw et al. (1994) (Fig. 7b).

The tracking of individual particles through the developing
thermal and physical structure of the orogen in this way
enables the calculation of model isotopic ages, as described in
Batt & Braun (1997). The distributions of modelled apatite
fission track, and biotite, muscovite and hornblende K–Ar ages
at the orogen’s surface after 10 Myr of deformation are shown
in Fig. 7(c). The depth within the deforming wedge at which
temperature exceeds that required for closure of the apatite
fission track system is so shallow that exposed material has
only cooled below that temperature within the last approxi-
mately 250 kyr, preventing the accumulation of a significant
quantity of fission tracks at uranium content as low as those
typically observed for apatite from the Alpine Schist in the
Southern Alps (Tippett & Kamp 1993). The modelled fission
track age of apatite is thus close to zero across much of the

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deformed wedge, increasing to between 0.5 and 2 Ma only within the bounding shear zones. Biotite and muscovite K–Ar ages are also low across the orogen, with consistent young ages observed from the retro-shear zone to some distance into the deformed wedge, at which point ages then increase rapidly with further distance towards the pro-side of the model. The distance at which this transition occurs is significantly greater for biotite ages than for muscovite. A similar distribution of ages is observed in the Southern Alps, where discrete zones of consistent young ages close to the Alpine Fault give way to older early Cenozoic to Mesozoic ages with distance to the southeast.

4.2 Model 2: influence of model rheology

In model 2 (M2), the upper 25 km of the crust now has a quartz-dominated rheology (R1), while the lower 5 km retains the feldspar-dominated rheology (R2) used in M1 (Fig. 5). This variation is introduced to reflect more closely the inferred crustal structure of the Pacific Plate in the central South Island, with 20–30 km of greywacke and schist overlying a layer of faster seismic velocity interpreted to be former oceanic crust (Reyners 1987; Reyners & Cowan 1993; Smith et al. 1995). With the basal temperature of the model fixed at 600 °C, the upper approximately 10 km of the quartz layer deforms entirely by frictional plastic behaviour (Fig. 8), while the lower part of the quartz layer and the underlying felsic layer are dominated to varying extents by thermally activated viscous processes.

Under conditions of active power-law creep, there exists a particularly high strength contrast at the contact between these two layers, where very weak quartzose material is juxtaposed against the relatively strong top of the felsic layer. Such a rheological contrast is consistent with the observed seismological character of the South Island, where strong seismogenic upper- and lower-crustal layers are separated by a relatively weak, aseismic mid-crust (Reyners 1987). In M2, the base of the upper-crustal quartzose layer is also significantly weaker than the base of the feldspar-dominated layer. This causes the two to decouple along their contact (Fig. 8), establishing an effective velocity singularity and divergence of material paths at the base of the quartz-dominated region. This decoupling brings material to the surface from the base of the quartz-dominated upper crust along the retro-shear zone of the model, while the lower, felsic-dominated layer is progressively thickened into a major growth fold structure at depth (Fig. 8). This growth fold pushes up the overlying crust as it develops, causing significant exhumation to the retro-side of the retro-shear zone. Due to the nature of the lower-crustal boundary preventing any crustal material from being subducted, this behaviour is not considered significant, except as regards the ultimate fate of the lower crust; the decoupling of the upper and lower crusts along the weak base of the quartzose layer make it very difficult to bring the felsic lower crust to the surface, even where it is here explicitly prohibited from subducting.

4.3 Model 3: surface processes

Where the strength of the deforming crust and the basal decoupling are low, as is the case with the upper quartzose crustal layer in M2, topography and the resulting gravitational
body forces have a significant influence on crustal deformation (Dahlen & Barr 1989; Koons 1989; Beaumont et al. 1992, 1996). This factor is investigated here in M3, where instead of being completely removed, material tectonically uplifted above the model’s base level is modified by a 1-D model of erosion and sedimentation processes.

The orographic influence of the Southern Alps is incorporated in this model by defining a finite moisture content for the atmosphere entering from the retro-side of the model. Above a basic background level of rainfall, precipitation in M3 is assumed to be proportional to surface elevation, but limited by the total amount of moisture present in the atmosphere. The parameter values used in this surface processes model were selected arbitrarily to produce a level of topography similar to that observed in the Southern Alps.

During the initial stages of orogen growth, development is identical to that expected for uniform precipitation (Fig. 9), with topography symmetrically distributed across the deformed region and growing gradually. By approximately 4 Myr into the experiment, however, sufficiently high topography has developed that the finite amount of precipitation entering the model from the retro- (or windward in this context) side is almost all captured by the retro-side of the orogen, while the pro-side is in an area of effective rain shadow, with precipitation limited to the imposed background level of the model. This immediately produces a strong asymmetry in orogenic development, as predicted by Koons (1990). Although the highest rates of uplift coincide with the retro-shear zone, the concentrated erosive power of the captured precipitation windward of the drainage divide maintains denudation at comparably high rates. In contrast, leeward of the drainage divide, the relatively low levels of erosion allow a significant proportion of the progressive surface uplift to accumulate, leading to rapid growth in topography (Fig. 9). Ultimately, the narrower retro-side of the orogen is dominated by the rapid erosion, which maintains slopes at a critical angle for the crustal material concerned, while the pro-side is wider, with lower slope angles developed by progressively accumulated uplift and lower erosion rates (Fig. 9). The maximum height that can be attained in this setting is limited by the amount of topography that can be built up in a given area leeward of the drainage divide before continued convergence takes that area into the rapid

Figure 8. Greyscale contour plot of the logarithm of strain rate after 10 Myr of tectonic deformation in model 2. Key to features is as outlined for Fig. 6(a), except that the bold black line represents the boundary between the upper and lower crust.

Figure 9. Topography developed in model 3. (a) Model topography at various times throughout the experiment. Each topographic profile is also offset by a vertical distance of approximately 1 km from the preceding one. (b) Variation in peak topography over time in model 3. Note the fundamental change in behaviour 4 Myr into the experiment. This corresponds to no physical change in model conditions, but rather represents the attainment of the critical topographic height to cause rainshadow conditions, as discussed in the text. Equivalent results for a model run with uniform precipitation are shown in grey for comparison.
eroded regime windward of the drainage divide. As the windward slope of the orogen is maintained at a material-dependent critical angle, this maximum height also limits the width of the windward zone (Fig. 9).

Although the timing of the different stages of this topographic development is arbitrarily determined by the parameter values selected, this model produces the observed asymmetrical dynamics and structure of the Southern Alps without the need for complexity in dynamical inputs. In particular, this simple model predicts a significant time lag between the initiation of rapid rock uplift and the development of rapid surface uplift and high topography such as seen in the Southern Alps today. This lag represents the time required to build up a rainfall barrier and produce the rainshadow effect and low erosion of the leeward side of the orogen responsible for the major build-up of topography. Such behaviour could account for the 2 Myr delay between the onset of rock uplift and the start of significant mean surface uplift inferred from fission track analysis of the Southern Alps east of the Main Divide (Tippett & Kamp 1995).

The topography developed in M4 and the resultant lithospheric flexure produce a relatively deep foreland basin close to the orogen on the pro-side of the model (Fig. 10b). This represents a notably poor match for the observed character of the Pacific Plate east of the Southern Alps in New Zealand, where the sedimentary veneer is a maximum of a few hundred metres thick beneath the Canterbury Plains onshore, and the maximum thickness of late Cenozoic sediment reaches only approximately 2.5 km, over 200 km from the Southern Alps in the Clipper Basin (Field et al. 1989). The results of modelling by Beaumont et al. (1996) suggest that different combinations of flexural and rheological parameters may be able to reduce this disparity, but no further effort is made here to improve the fit of basin development in these models to observations from the Southern Alps.

4.4 Models 4, 5 and 6: tectonic evolution

It is generally agreed that the Southern Alps owe their present existence to a Neogene reorganization of relative Pacific–Australian Plate motions (Walcott 1984). Exactly when and how rapidly this change occurred, however, has been a matter of recent debate. Based on a simple interpretation of thermochronology from the Mount Cook region, Tippett & Kamp (1993) suggested that the recent rapid convergence seen in the South Island has only developed in the past 1.3 ± 0.3 Ma (see Fig. 15 in Tippett & Kamp 1993), with convergence prior to this at much lower rates. In contrast, from close examination of tectonic markers in the broader southwest Pacific region and the sedimentary record from around New Zealand, Sutherland (1995, 1996) proposed that this change occurred in a single rapid reorganization of plate motion at approximately 5 Ma. Along different lines again, Shi et al. (1996) modelled the development of the Southern Alps by assuming slow convergence prior to 5 Myr ago, and then a linear increase in convergence rate between 5 Ma and the present.

These hypotheses each start with the constraint offered by the present state of the orogen, but differ significantly in their interpretation of its early history. With the rapid uplift and erosion of the Southern Alps, the current nature of the orogen gives little insight into this early development. The distribution of isotopic ages, however, gives a direct record of the thermal development of the Southern Alps up to 15 Myr into the past (Fig. 4). Where these ages can be placed firmly in a dynamical context, as is possible here through our coupled modelling approach, they provide a test of the relative merits of these proposed histories. Due to the complex, multivariate nature of both the model used here and the process of open-system behaviour in thermochronological systems (e.g. Sardarov 1957; Hurley et al. 1962; Dodson 1973), there is little to be gained from trying to match the ages observed for a region in detail. Rather, the approach taken here is to assume that the many variable factors controlling this process are relatively consistent within the Southern Alps, and thus to look for some aspect of the model that can produce the style of variation observed in the thermochronological data across the orogen.

Models 4, 5 and 6 explicitly test the thermal and thermochronological consequences of each of the hypotheses outlined above. These models are physically identical to M3, and each is run for 10 Myr. In M4, convergence velocity is constant at 2 mm yr⁻¹ for the initial 5 Myr, and then increases to 10 mm yr⁻¹ thereafter, as in the hypothesis of Sutherland (1995). M5 represents the evolutionary style suggested by Shi et al. (1996), with convergence of 2 mm yr⁻¹ for the initial

Figure 10. Greyscale contour plot of the logarithm of strain rate in model 4 after 10 Myr of deformation. Key to features is as outlined for Fig. 8, except that the thin black lines in the upper left and right corners of the model represent the sedimentary stratigraphy of the evolving foreland basins.

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5 Myr, and then a linear increase in convergence rate up to a maximum of 10 mm yr\(^{-1}\) at the end of the experiment. M6 is designed to reflect the hypothesis of Tippett & Kamp (1993), with convergence constant at 2 mm yr\(^{-1}\) between 10 and 1.3 Ma and thereafter increasing to 10 mm yr\(^{-1}\). It should also be noted that although these histories give varying levels of total convergence—60 km in M4, 40 km in M5 and 30.4 km in M6—these all fall within the range of current estimates for this parameter of 50 ± 35 km (Walcott 1984).

After the first 5 Myr of slow convergence in M4, the initially diffuse pro- and retro-shear zones immediately sharpen and then progressively develop an asymmetry, with the retro-shear zone becoming far more prominent than the pro-shear zone due to the perturbation of the geothermal structure of the orogen, as described for M1. The developing thermal perturbation also causes the initiation of a prominent secondary pro-shear zone within the deforming wedge.

The slow uplift of the first 5 Myr maintains the topography of M4 at a low level, approaching dynamic equilibrium between uplift and erosion towards the end of this period (Fig. 11). With the increase in convergence rate to 10 mm yr\(^{-1}\), topography initially increases sharply before again moving towards a steady-state level (Fig. 11). At around 7 Myr into the experiment, however, the model reaches the topographic height required for it to capture orographically most of the rain coming from the retro-side of the model, leading to extremely rapid topographic growth and asymmetric orogenic development as outlined for M3. Again, it should be stressed that the heights and timings of these different stages in topographic development are arbitrary, but that this progression illustrates the possibility of a significant delay between an increase in convergence rate and accompanying change in topographic style.

Fig. 12(a) plots the minimum apparent ages observed at the surface of M4 at each time step. Following the increase in convergence velocity at 5 Myr, this minimum age value evolves through three stages for any given thermochronometer (Fig. 12a). Ages initially decrease linearly with time (stage 1), then fall sharply over a brief interval (stage 2), after which they maintain consistent low levels for the remainder of the experiment (stage 3). Stages 1 and 2 correspond to adjustment between the thermal regime of the model at 5 Ma and that eventually attained following the increase in convergence rate after 5 Ma. The isotopic ages observed at the surface of an exhumed metamorphic belt relate to material cooling through the closure temperature of the chronometer concerned at some finite depth within the deforming crust. An age observed at the surface thus represents the total time taken to exhume material from the point at which it initially cooled through the relevant closure isotherm.

The linear ‘stage 1’ decrease in minimum age observed over the last 5 Myr of M5 marks the progressive exhumation of a layer of material which had already cooled to below the relevant closure temperature prior to the change of velocities at 5 Ma. The relative proportion of the exhumation history of exposed material spent in the post-5 Ma regime of rapid exhumation increases with time after the increase in convergence velocity. This results in the net time taken to exhume material following closure, and thus its apparent age when eventually exposed at the surface decreasing over this interval.

An increase in the rate of exhumation also causes re-organization of geothermal structure, with isotherms advected to shallower depths and the near-surface geothermal gradient increased. The depth at which material cools through a given closure temperature, and thus the time taken to expose material at the surface following closure, decreases progressively during such reorganization of the geothermal structure. The later, more rapid drop in age (stage 2) reflects the exposure of such material, which was on the verge of cooling through the relevant closure temperature at 5 Ma, and thus does so as this thermal reorganization is occurring in the early stages of the regime of rapid uplift and exhumation in the final 5 Myr of the experiment. The active advection of the isotherms adds to the effects described for stage 1 above, giving the observed sharp drop in ages (Fig. 12). Following this period of re-organization, the geothermal structure of the model stabilizes in a new steady-state configuration, leading to the later consistent ages (stage 3) observed in M4.

Because the transition from stages 1–2 occurs due to the thermal perturbation associated with the increase in convergence velocity and the attendant exhumation rate, the age at which each chronometer in M4 displays the transition from stage 1 to stage 2 ages corresponds to the time elapsed since 5 Ma (Fig. 12a). Where such a transition is apparently observed for the K–Ar age of micas in the Fox Glacier region of the modern Southern Alps (Fig. 4), it suggests an age of approximately 5 Myr (Fig. 4) for the inferred episode of geothermal reorganization (Batt 1997), in agreement with the hypothesis of Sutherland (1995).

M5 is identical to M4 for the first 5 Myr of the experiment (Fig. 5). After this initial period, the progressive velocity increase in M5 leads to a gradual intensification of the pro- and retro-shear zones. These shear zones are always more diffuse and less asymmetrical than those observed in M4 at an equivalent time, due to the lower exhumation rate and consequently less intense thermal perturbation of M5 compared to M4. These structural differences between M4 and M5 lessen as M5 approaches a convergence velocity of 10 mm yr\(^{-1}\), with the strain-rate distribution of the two almost identical by the end of the respective experiments.

![Figure 11. Peak topographic height over time relative to model base level in model 4. Note the division into three distinct behavioural zones.](https://academic.oup.com/gji/article-abstract/136/2/403/694251)
Despite this convergence in structural character over time, the isotopic age record of M5 is significantly different from that described for M4. The lower total exhumation of M5 exposes reset K–Ar ages for biotite and muscovite later than in M4 (Fig. 12c), with exhumation not reaching to depths at which amphibole is experiencing diffusive loss of argon at all during M5. For muscovite, biotite and apatite, the progressive increase in convergence velocity (and hence exhumation rate) with time in M5 masks any transitional (stage 2) age effects caused by advection of the relevant isotherms. Notably, despite these differences, the similarity in the histories of models 4 and 5 towards the end of the respective experiments is sufficient to give the two indistinguishable apatite fission track ages after 10 Myr (Figs 12a and c).

The initial 5 Myr of the development of M6 is again identical to that described for M5 (Fig. 5). This pattern of strain accommodation is maintained in an effective steady state in M6 until the sudden increase in convergence velocity experienced 8.7 Myr into the model. For the ensuing 1.3 Myr, the development of M6 is then as outlined for the early period of adjustment to rapid convergence velocity for M4.

As with M4, the sudden change in convergence rate produces a related transitional (stage 2) age pattern caused by the rapid perturbation of the geothermal structure (see apatite fission track curve in Fig. 12d). In M6, however, these stage 2 patterns occur 1.3 Myr before the end of the experiment, rather than the 5 Myr apparently observed in the record of the Southern Alps. There is also a lack of sufficient exhumation in M6 to see these effects expressed for thermochronometers other than fission tracks in apatite.

Despite these differences, the convergence of dynamics towards the present day in models 4 and 6 again gives two virtually identical apatite fission track ages at the ends of the respective experiments (Fig. 12). This convergence of ages is due to the low annealing temperature of fission tracks in apatite (Brandon et al. 1998), and the consequent shallow depth of exhumation required to ‘reset’ the thermochronological record for that system following a change in dynamics. As these results amply demonstrate, while fission tracks provide the best indication of the recent dynamical behaviour of an eroding orogen, the early history of the deeply exhumed parts of the Southern Alps cannot be fully constrained by this low-temperature method of thermochronology. In this respect, thermochronometers with higher closure temperatures provide significantly better tests of alternative dynamical histories that converge to the present day, such as those examined in this study.

### 4.5 Model 7: variation in convergence velocity along the strike of the orogen

The convergent plate motion responsible for the uplift of the Southern Alps is due to the obliquity of relative Australian–Pacific Plate motion to the boundary between these two plates through the South Island (Walcott 1978, 1984; Norris et al. 1990). According to the Global Positioning System measurements of Smith et al. (1996), which place the present instantaneous Euler pole for motion between these plates at $62.9 \pm 1.0 \degree S$, $179.9 \pm 0.4 \degree W$, the orientation of the Australian–Pacific Plate relative velocity vector should become closer to...
Parallel to the orientation of the Alpine Fault (thus decreasing the convergent component of the plate motion) from north to south along this boundary (Table 2). This relationship breaks down in the north of the orogen, where the Marlborough Fault System and related structures (Fig. 1) accommodate a significant proportion of the relative Australian–Pacific Plate motion as strike-slip displacement (Allis 1986; Kamp et al. 1989).

M7 seeks to determine whether the magnitude of this predicted variation in convergence velocity is sufficient to explain the systematic variation in isotopic ages and other constraints along the Southern Alps. This model is identical to M4 except for the convergence velocity experienced in the final 5 Myr of the experiment, which varies from 5 to 15 mm yr\(^{-1}\) in increments of 1 mm yr\(^{-1}\) (Fig. 5).

The basic evolution of these models following the increase in convergence velocity 5 Myr into the experiment is largely consistent, with all bar one developing an asymmetric strain rate distribution with prominent retro-shear and relatively weak, distributed primary pro-shear zones. Only the slowest model differs significantly from this pattern, having a more prominent pro-shear zone, and thus a more symmetrical character than do the others. This symmetry may be due to a relative absence of thermal perturbation and a consequent lack of weakening of the retro-shear zone in this model. All of this model series bar the slowest also develop a discrete secondary pro-shear zone (as described for M1) which stabilizes with its surface trace on the pro-side of the drainage divide.

At the parameter values used for surface processes in this experiment, the slowest of the M7 series attains steady-state topography at a level below the height required to form an effective orographic rainfall barrier (Fig. 13), but this critical height is reached by all the other models in this series. Greater velocity of convergence results in progressively earlier attainment of the topographic limit and the resultant development of rapid asymmetric topographic growth, as described for M3.

The variation in convergence rate between the various M7 experiments gives a variation in total exhumation (and hence in the apparent ages observed at the surface of the model) comparable to that observed through time in model 4 (Fig. 14). For models with low convergence velocity, and thus little total exhumation, the ages of a given chronometer decrease sharply with increasing exhumation (Fig. 14), reflecting the exposure of material that had already cooled through the relevant closure temperature prior to 5 Ma. At greater exhumation levels, there is then a transition to younger ages which reflect the perturbed geothermal structure of the recent regime of rapid convergence and exhumation. Unlike in M4, however, where ages in this young zone remain at a consistent level, the higher velocities giving increasing exhumation through the M7 series also perturb the geothermal gradient of the model to a greater extent. As a result, the young (stage 3) ages observed at the end of the experiment (10 Ma) progressively decrease as convergence velocity increases (Fig. 14). This style of variation is a better fit than that of M4 to the observed isotopic ages of the Southern Alps, which show systematic variation along the orogen, even in the region of low, relatively consistent ages between the Copland and Whataroa Valleys (Fig. 4). This suggests that differences in exhumation rate (related in this model to variation in the rate of convergence experienced) are a more likely source of the observed character of the Southern Alps than diachronieity in orogenic processes.

### 5 DISCUSSION AND CONCLUSIONS

As suggested previously by Allis et al. (1979), Koons (1987) and Shi et al. (1996), the rapid uplift and exhumation experienced in the Southern Alps are responsible for significant perturbation of the geothermal structure beneath the orogen. The modelling in this study suggests that this in turn may have had a significant influence on the physical development of the orogen. Advection of hot lower-crustal material along the retro-shear zone of an orogen (the Alpine Fault in the case of the Southern Alps) thermally weakens this shear zone and leads to asymmetry of development in the orogen, with the retro-shear zone becoming far more prominent than the pro-shear zone as a region of high strain rate. This process could...
account for the prominence of the Alpine Fault as a discrete feature in geophysical analyses of the Southern Alps (Smith et al. 1985; Davey et al. 1995). As noted by Allis et al. (1979) and Koons (1987), this thermal weakening would also give an overall thinning of the seismogenic zone in the area of the retro-shear, offering one possible explanation for the low seismicity of the Alpine Fault in central areas of the South Island (Walcott 1978; Allis et al. 1979). In our present study, such thermal weakening is also shown to lead to the formation of a prominent secondary pro-shear zone in the upper crust that offers an explanation for the presence of the Main Divide Fault Zone in the Southern Alps.

The significance of this thermal weakening is in conflict with the modelling of Shi et al. (1996), who suggest that the brittle–ductile transition may actually be deflected downwards beneath the Southern Alps. The models used by Shi et al. (1996), however, do not incorporate temperature-sensitive rheologies, with the inferred ‘brittle–ductile transition’ based solely on the position of the 350°C isotherm. The downward deflection of this boundary is in turn linked to an arbitrary balance between the amount of material uplifted to form the Southern Alps and the amount of thickening at depth to produce a crustal root (Shi et al. 1996). As the uplift of the brittle–ductile transition in the rheologically sensitive models used here explicitly accounts for many important physical features of the Southern Alps without producing thermal or thermochronological results significantly different from the predictions of Shi et al. (1996), it is suggested that the present study offers a more reasonable view of the thermomechanical structure of the orogen.

As previously suggested (e.g. Koons 1990; Beaumont et al. 1992), asymmetry of surface processes is also highly significant in the physical development of the Southern Alps, potentially explaining the highly asymmetric structural and topographic nature of the orogen and the inferred delay between the initiation of rock uplift and the development of significant surface uplift in the Southern Alps.

The details of variation in age along the orogen strongly support the suggestion of Sutherland (1995, 1996) that the present tectonic regime of the Southern Alps developed in a single rapid reorganization of Australian-Pacific relative plate motions at approximately 5 Ma and has remained stable since that time. Models incorporating this simple two-stage dynamical history produce model thermochronological records comparable to those observed in the Southern Alps, where distinctly different trends in K–Ar mica ages are observed in southern and central regions of the orogen.

That different stages of this thermochronological development are exposed at different points along the Southern Alps can be explained by a progressive reduction in the total exhumation experienced over the past 5 Myr from the Whataroa River southwards along the orogen. This is consistent with the variation in the velocity of convergence between the Pacific and Australian plates predicted from the GPS measurements of Smith et al. (1996).

The insights gained into the development of the orogen also demonstrate the power that thermochronology, as a direct record of thermal development, offers as a constraint on a region’s tectonic evolution. This is especially true of the earlier stages of orogenic development that may no longer be preserved in discrete physical features at the surface.

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