Construction of a composite pressure–temperature path: revealing the synorogenic burial and exhumation history of the Sevier hinterland, USA

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ABSTRACT Metamorphic pressure–temperature (P–T) paths derived from 16 growth-zoned garnets, nine from this study and seven from a previous study, have been combined to construct a detailed composite path for an area in the hinterland of the Cretaceous to early Tertiary Sevier orogenic belt in southern Idaho and north-west Utah. Samples are from two Proterozoic units in the footwall of the Basin-Elba thrust: the schist of Mahogany Peaks in the central Albion Mountains, Idaho, and the schist of Stevens Spring in the Basin Creek area of the Grouse Creek Mountains, Utah, ∼40 km to the south. The simulated portions of the garnets analysed in this study grew from reactions involving the breakdown of chlorite in the upper greenschist to lower amphibolite facies. Multiple garnets were analysed from three samples. Overlapping segments of P–T paths from different garnets in the same sample correlate with respect to slope and garnet Mn concentration. The composite P–T path documents three episodes of sharply increasing pressures separated by two episodes of pressure decrease, all during progressively increasing temperatures. The path is interpreted to represent alternating episodes of synconvergent thrusting and extensional exhumation in the hinterland of the Sevier orogen. Burial was probably caused by the Basin-Elba fault, the only major thrust exposed in the region. Extensional exhumation may have occurred along the Mahogany Peaks or Emigrant Spring faults, or by extensional reactivation of the Basin-Elba fault. This method of correlating partial P–T paths to reveal a more complete composite path provides a powerful tool in unveiling orogenic histories in metamorphic terranes, where evidence of major structures responsible for burial and exhumation is commonly obscured by later events.

Key words: garnet; Gibbs method; P–T path; Sevier; Utah.

INTRODUCTION

Metamorphic rocks preserved in ancient mountain belts represent one of the primary and potentially enduring records of past orogenesis. The cycle of orogenic activity experienced by mountain belts commonly culminates with profound extension and magmatism (Dewey, 1988) which can obscure earlier structures and fabrics formed during shortening. Consequently, the reconstruction of changes in pressure and temperature from metamorphic rocks provides a view of past tectonism not always apparent from other geological observations (e.g. Selverstone, 1985; Spear et al., 1990).

Tectonic events including subduction, continental collision, extension, lower lithospheric delamination and arc magmatism impart characteristic changes in pressure and temperature to rocks during metamorphism; these changes in pressure and temperature (P–T paths) can be linked to tectonic processes through increasingly sophisticated thermal modelling (Oxburgh & Turcotte, 1974; England & Thompson, 1984; Ruppel et al., 1988; Grasemann & Mancktelow, 1993; Platt & England, 1994; Ruppel & Hodges, 1994; Ketcham, 1996; Jamieson et al., 1998, 2002; Hoisch, 2005). Despite their great utility, the reconstruction of complete P–T paths remains one of the foremost challenges in metamorphic petrology (Spear, 1993).

The construction of complete P–T paths is hindered by a number of factors including: (i) lack of development or preservation of mineral assemblages appropriate for analysing P–T paths during all segments of the path, (ii) erasure of earlier histories through partial or complete diffusive re-equilibration if temperatures exceed thresholds for cation diffusion, and (iii) difficulty in correlating P–T paths from rocks recording different or overlapping segments of paths from the same locality. Although no single method exists to accurately determine all parts of a P–T path in a given area, from the earliest burial to the final exhumation, methods which utilize growth zoning in garnet, such as the Gibbs method based on Duhem’s theorem (e.g. Spear, 1988), offer the greatest potential to record the most complete P–T path. Metamorphic garnet will record the portions of the prograde path through which they grew, as long as temperatures do not exceed thresholds for cation volume diffusion.
(600–650 °C, depending on cooling rate, e.g. Spear, 1989).

Orogenic $P$–$T$ paths have classically been thought to have the form of a simple clockwise loop, reflecting a basic concept of orogenesis starting with protracted shortening, followed by thermal relaxation, and terminating with exhumation by combined erosion and extension (e.g. England & Thompson, 1984). In parallel with the recognition that plate boundary forces, gravitational potential energy, rock rheology and erosion dynamically interact to influence deformation kinematics (Davis et al., 1983; Platt, 1986; Molnar & Lyon-Caen, 1988; England & Houseman, 1989; Royden, 1993; Rey et al., 2001), orogenic histories involving synconvergent extension and alternations in shortening and extension have been documented (Hodges et al., 1992b; Wallis et al., 1993; Wells, 1997; Ferranti & Oldow, 1999; Collins, 2002). Despite the theoretical predictions and field examples of alternating shortening and extension, $P$–$T$ paths confirming such alternations are not widely reported (e.g. Hoisch et al., 2002; Johnson & Brown, 2004).

In this study we present an unusually complete composite metamorphic $P$–$T$ path determined from exhumed mid-crustal pelitic schists from the Sevier orogenic belt of the western United States. The garnets analysed in this study nucleated at different times and thus captured different segments of the path. The $P$–$T$ paths recorded in individual garnets overlap and were correlated to construct a composite $P$–$T$ path. The composite path is much more complex than the simple $P$–$T$ loops commonly assumed for tectonic burial and uplift, and is interpreted to represent alternating episodes of synconvergent thrusting and extensional exhumation.

**REGIONAL TECTONIC HISTORY**

The metamorphic rocks in this study are from the hinterland of the Late Mesozoic to Early Cenozoic Sevier orogen of the western United States. The Sevier orogen represents a classic retroarc orogenic wedge, broadly similar in gross architecture to the modern Andean orogen (Allmendinger, 1992; Burchfiel et al., 1992; Camilleri et al., 1997; DeCelles, 2004). The Sevier orogen is subdivided into an eastern foreland fold-thrust belt and a western hinterland region that includes the isolated exposures of greenschist and amphibolite facies Barrovian metamorphic rocks of the Cordilleran metamorphic core complexes (Crittenden et al., 1980). Initial shortening in the Sevier orogen progressed eastward from hinterland to foreland from Middle Jurassic to Early Eocene time (Armstrong & Oriel, 1965; Wiltshko & Dorr, 1983; Elison, 1991; Allmendinger, 1992; Taylor et al., 2000; DeCelles, 2004). The progressive development of the foreland fold-thrust belt from Early Cretaceous to Early Eocene is well understood, with an eastward migrating thrust front punctuated by local minor out-of-sequence deformation internal to the orogenic wedge (DeCelles, 1994; Lawton et al., 1997; Yonkee et al., 1997; DeCelles, 2004).

The kinematic history of the hinterland is less clear and includes a sparse record of Mid-Late Jurassic shortening (Allmendinger et al., 1984; Elison, 1991; Miller & Allmendinger, 1991; Hudec, 1992) as well as Late Cretaceous shortening and extension (Miller et al., 1988; Hodges & Walker, 1992; Camilleri & Chamberlain, 1997; Wells, 1997; Wells et al., 1998; McGrew et al., 2000). Original stratigraphic juxtapositions and deformation fabrics recording early shortening are commonly obscured by reactivation and overprinting by normal faults and extensional flow fabrics (Camilleri & Chamberlain, 1997; Wells, 1997; Lewis et al., 1999). The metamorphism provides the strongest evidence for large magnitude shortening in the hinterland of the orogen as: (i) systematic lateral gradients in metamorphic pressure in stratigraphically equivalent and contiguous metamorphic rocks provide evidence for lateral gradients in thrust loading (Hoisch & Simpson, 1993; Camilleri & Chamberlain, 1997; Lewis et al., 1999); (ii) Barrovian metamorphism of miogeoclinal metasedimentary rocks to pressures as high as 8–10 kbar indicates structural burial of two to three times that of stratigraphic depth (e.g. Hodges et al., 1992a; McGrew et al., 2000; Hoisch et al., 2002).

**GEOLOGY OF THE RAFT RIVER-ALBION-GROUSE CREEK MOUNTAINS**

In the Raft River, Albion and Grouse Creek mountains of north-west Utah and southern Idaho (Fig. 1), a succession of tectonically-thinned greenschist to upper amphibolite facies Mississippian to Proterozoic metasedimentary rocks, termed the Raft River Mountains sequence by Miller (1983), unconformably overlies Archean basement (Armstrong, 1968; Compton et al., 1977; Wells, 1997). This succession occurs over an area greater than 4000 km² in all parts of the region except in the hangingwall of the Basin-Elba thrust fault in the northern Albion Mountains (Miller, 1980, 1983), which is composed of a different sequence of rocks referred to by Miller (1983) as the ‘quartzite assemblage’, a >3500 m thick succession of overturned Late Cambrian and Neoproterozoic quartzite and schist (Mt. Harrison sequence) overlain tectonically by a sequence of upright quartzite and schist of inferred Late Cambrian and Neoproterozoic age (Robinson Creek sequence). The Basin-Elba fault is apparently the only preserved thrust fault in the region as it places older strata over younger (Figs 1 & 2).

Exhumation of the metamorphic rocks took place during episodic Cenozoic extension along both west and east vergent detachment fault systems (Compton et al., 1977; Malavieille, 1987; Wells et al., 2000b;
Fig. 1. Generalized geological map of the Raft River-Albion-Grouse Creek Mountains, showing location of the Basin-Elba fault, Basin Creek area, and approximate sample locations. Bottom right inset shows location of metamorphic core complexes (solid areas) in the hinterland of the Sevier thrust belt and in the southern Basin and Range. The thick barbed line represents the leading edge of the thrust belt. RCS, Robinson Creek sequence; MHS, Mount Harrison sequence (Miller, 1983). Modified from Hoisch et al. (2002).
Egger et al., 2003); synconvergent Mesozoic extension was also important (e.g. Wells et al., 1990, 1998; Hodges & Walker, 1992). The oldest and most widely distributed extensional fabric (Table 1) is a flat-lying foliation and generally N-trending stretching lineation interpreted to have developed during orogen-parallel extensional flow at 105 ± 6 Ma (revised from Wells et al., 2000a). This fabric affects Archean to Permian rocks and accomplished variable but significant thinning of rock units. Additionally, low-angle, generally bedding-parallel, top-to-west faults, including the Mahogany Peaks and the Emigrant Spring faults (Fig. 2c), accomplished significant attenuation of the rock units and are interpreted as extensional faults (Table 1) (Wells, 1997; Wells et al., 1998). The Mahogany Peaks fault juxtaposes Ordovician marble over the Proterozoic schist of Mahogany Peaks with an estimated excision of 4–5 km, and its age is bracketed by 40Ar/39Ar muscovite cooling ages of c. 90 and 60 Ma from the hangingwall and footwall respectively (Wells et al., 1998). The Emigrant Spring fault (Wells, 1997) places Pennsylvanian (?) calcitic marble over Ordovician and Silurian (?) dolomitic marble and removes ~5 km of stratigraphic section. Motion on the Emigrant Spring fault has been previously interpreted as prior to or synchronous with cooling recorded by an 40Ar/39Ar muscovite age of c. 89–88 Ma from a greenschist facies calcitic marble mylonite (Wells et al., 1990; Wells, 1997); however, the cooling age may pre-date deformation as the temperatures of deformation may have been less than argon closure in muscovite. Kilometre-scale recumbent folds deform the Emigrant Spring and Mahogany Peaks faults, leading to an interpretation of renewed contraction following extension (Wells, 1997; D3 in Table 1). These recumbent folds are overprinted by an Eocene–Oligocene extensional shear zone and detachment fault (D4 in Table 1).

The P–T paths generated in this study are from two garnet-bearing units of Proterozoic schist in the footwall of the Basin-Elba fault. The schist of Mahogany Peaks was sampled close to the Basin-Elba fault in the Albion Mountains (Figs 1 & 2a), and the ‘upper horizon’ of the schist of Stevens Spring was sampled from the Basin Creek area of the Grouse Creek Mountains, ~40 km to the south (Figs 1 & 2b). As these two localities lie within a contiguous exposure of the Raft River Mountains sequence and no bedding parallel faults are recognized between them, the two units are considered to have shared a similar P–T history.

Hoisch et al. (2002) reported P–T paths from the Basin Creek area from two different horizons of the schist of Stevens Spring, a ‘lower horizon’ and an ‘upper horizon’. The upper horizon garnets grew at ~475 °C and records one episode of approximately isothermal pressure increase, interpreted to represent rapid burial associated with thrusting. Garnet in the lower horizon grew at higher temperatures (600–630 °C) from a different reaction, and record a second episode of burial following an episode of major decompression (Hoisch et al., 2002). In this study, five new P–T paths are presented from two localities of the schist of Mahogany Peaks, and four new P–T paths are presented from the upper horizon of the schist of Stevens Spring. These are integrated with the seven paths previously determined from the schist of Stevens Spring (five from the upper horizon and two from the lower horizon) (Hoisch et al., 2002) (Fig. 2).

ANALYTICAL METHODS

Mineral compositions were determined using the Cameca MBX electron microprobe at Northern Arizona University. A spot size of 1 μm and a beam current of 25 nA was used for garnet traverses. A 5 μm spot size and a beam current of 10 nA was used to acquire matrix mineral data. The accelerating voltage in both cases was 15 kV. Line traverses across garnet were run with a point spacing of 10–20 μm (Fig. 3). The garnet traverse data were processed to exclude any analysis that failed to produce a good stoichiometry, within certain tolerances. Generally, excluded analyses represented inclusions or points that overlapped

Table 1. Sequence of Mesozoic to early Cenozoic deformation proposed in this and previous studies (see text) for the footwall to the Basin-Elba thrust fault, western Raft River, northern Grouse Creek and Albion Mountains.

<table>
<thead>
<tr>
<th>Tectonic event</th>
<th>Structure and kinematics</th>
<th>Interpretation</th>
<th>Timing</th>
<th>Metamorphism</th>
</tr>
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<tbody>
<tr>
<td>M1</td>
<td>First motion on Basin-Elba fault</td>
<td>Contraction</td>
<td>Pre-105 Ma, Jurassic or early Cretaceous</td>
<td>Metamorphism of Raft River assemblage; isothermal P increase&lt;sup&gt;b&lt;/sup&gt;</td>
</tr>
<tr>
<td>D1</td>
<td>Top-to-north shearing&lt;sup&gt;a&lt;/sup&gt;</td>
<td>Orogen-parallel extension&lt;sup&gt;c&lt;/sup&gt;</td>
<td>105 ± 12 Ma&lt;sup&gt;ab&lt;/sup&gt;</td>
<td>Inferred decompression P–T path&lt;sup&gt;b&lt;/sup&gt;</td>
</tr>
<tr>
<td>D2</td>
<td>Emigrant Spring and Mahogany Peaks Fault</td>
<td>Extension&lt;sup&gt;c&lt;/sup&gt;</td>
<td>c. 90–60 Ma&lt;sup&gt;a&lt;/sup&gt;</td>
<td>P increase&lt;sup&gt;b&lt;/sup&gt;</td>
</tr>
<tr>
<td>D3</td>
<td>Recumbent folding and second motion along Basin-Elba fault</td>
<td>Contraction&lt;sup&gt;c&lt;/sup&gt;</td>
<td>c. 60–45 Ma&lt;sup&gt;a&lt;/sup&gt;</td>
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<tr>
<td>D4A</td>
<td>Top-to-WNW shearing along amphibolite facies MMSZ</td>
<td>Extension</td>
<td>Late Eocene&lt;sup&gt;a&lt;/sup&gt;</td>
<td></td>
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<sup>a</sup>This study.
<sup>b</sup>Hoisch et al. (2002).
<sup>c</sup>Wells et al. (in press).
<sup>d</sup>Malavieille (1987).
<sup>e</sup>Wells (1997).
<sup>f</sup>Wells et al. (1998).
<sup>g</sup>Revised from Hoisch & Wells (2004).
<sup>h</sup>Wells et al. (2000b).
Fig. 2. Geological maps of the sampled areas (see Fig. 1) and regional stratigraphy. (a) Map of the northern Albion Mountains (based on Miller, 1983 and R.L. Armstrong, unpublished mapping) and locations for sample sites THAL4 and THAL6. (b) Map of the Basin Creek area of the northern Grouse Creek Mountains (modified from Hoisch et al., 2002) and locations for sample sites UH and LH. All map units are keyed to the tectonostratigraphic column (c), with the exception of Oligocene granite (black fill).
inclusions. Multiple analyses of each matrix phase were performed in each sample and averaged (Table S1). Garnet element maps for Ca, Mg, Fe and Mn were generated by counting with wave dispersive spectrometers on a JEOL-733 electron microprobe at Rensselaer Polytechnic Institute and by counting with an energy dispersive system on a JEOL JSM-6480LV scanning electron microscope at Northern Arizona University (Fig. 4). Images were processed with NIH IMAGE or IMAGE J to produce maximum contrast.

**NUMERICAL SIMULATION OF GARNET GROWTH USING THE GIBBS METHOD WITH DUHEM’S THEOREM**

Pressure–temperature paths were determined from garnet growth zoning by numerically simulating garnet growth based on the Gibbs method with Duhem’s theorem (Spear, 1988, 1993). Garnet in this study display bell-shaped Mn profiles, consistent with Rayleigh fractionation during equilibrium growth (e.g. Hollister, 1966). Zoning is generally concentric except for post-growth modifications (Figs 3 & 4). The zoning profiles range from highly symmetric to slightly asymmetric.

Garnet growth is accompanied by changes in intensive variables (composition of solid solution phases, pressure and temperature) and extensive variables (amounts of phases). According to Duhem’s theorem, chemical systems possess two degrees of freedom. By specifying the changes in any two variables (the ‘monitors’ parameters’), the changes in all the others can be solved. The program GIBBS (Spear et al., 1991), version Feb. 12, 1999, was used to perform the calculations. Items required for setting up the calculation include specifying the nucleation density (number of garnet nuclei per 100 cm$^3$ of rock), the phases in equilibrium with garnet during its growth, the initial compositions of the solid solution phases including garnet, the modal abundances of the phases at the time garnet begins growing, the temperature and pressure corresponding to initiation of garnet growth, the choice of monitors and the values of the monitors. Results obtained by this method are statistically robust (Kohn, 1993).

For all garnet growth simulations in this study, a model system appropriate to pelitic schist was assumed: K$_2$O-Na$_2$O-CaO-Al$_2$O$_3$-SiO$_2$-MgO-FeO-MnO-H$_2$O. The monitors were selected to be the grossular content of garnet ($X_{gr}$) and the amount (mole) of garnet ($M_{Gar}$) in 100 cm$^3$ of rock. The grossular content of garnet was selected because it has been shown to be the least susceptible of the eight-fold cations to diffusion (e.g. Spear, 1993). The changes in grossular content ($\Delta X_{gr}$) during garnet growth may be calculated from data contained in the zoning profile, and the changes in the amount of garnet is determined through trial and error, that is, when the simulation has produced garnet of the correct radius, then the correct value of $\Delta M_{Gar}$ has been entered. The calculation solves for all remaining extensive and intensive parameters, including pressure and temperature, from which the $P$–$T$ paths are constructed. For all simulations in this study, the ideal one-site mixing activity models pre-specified in the GIBBS program were used for staurolite (Fe, Mg, Mn), garnet (Fe, Mg, Ca, Mn), chlorite (Fe, Mg, Mn), paragonite (Ca-K-Na) and plagioclase (Ca-Na), and pure end-members with activities of 1.0 were for assumed for muscovite and quartz. The thermodynamic data set used is contained in the Thermo.Dat file included with the GIBBS program, which is described in Spear & Cheney (1989, their table 2) for the minerals of interest.

**Initial compositions and modes**

Of all the initial conditions required to be specified, only the initial composition of the garnet, corresponding to the core, is directly measurable. In most simulations, the actual (measured) compositions of the other minerals were used as initial values (Table S2) because the simulations predicted only slight changes in composition and mode during garnet growth. In all cases in this study, garnet was determined to have grown through chlorite-breakdown reactions (discussed later on for each case), and in all cases, chlorite was eliminated from the actual rock following garnet growth. Consequently, both the initial mode and initial composition of chlorite had to be estimated. In general, the initial mode of chlorite in the simulation was assumed to be ~1.5 times that of the actual garnet mode. The chlorite initial Fe/Mg ratio was estimated by back calculating using the garnet-chlorite thermometer of Grambling (1990), the assumed initial temperature for garnet growth, and the measured garnet core composition (Table S1). A chlorite stoichiometry typical of pelitic schist ($R^{2+}$, Al, Si in the proportions 4.5:3:2.5) was assumed, which is a pre-specified option in the GIBBS program.

**Initial temperature and pressure**

The initial temperature and pressure assumed in the models were determined through a variety of methods, as described later on for each sample. Mineral assemblages in relation to known stability constraints and thermobarometry were useful in estimating initial conditions, but in all cases in this study, the determinations carry large uncertainties.

**Nucleation density**

The values assigned to nucleation density act upon the calculation to divide the garnet grown into a specified number of nuclei. Increasing the nucleation density results in smaller garnet, all other things being equal. This gives the user a second way to adjust the garnet
radius in the simulations, the other being altering the value of $D_{M,Gar}$. However, the two methods are not equivalent, as altering the value of $D_{M,Gar}$ will modify the extent of reaction, whereas changing only the nucleation density will not.

**Mineral assemblages**

In all cases in this study, the mineral assemblage present at the time of garnet growth was determined to be different than the mineral assemblage of the actual rock. Thus, it was necessary to determine the full reaction history of each sample so that the correct minerals could be associated with garnet growth in the simulations. Reaction histories are discussed for each sample in the subsequent sections. All garnets simulated in this study are interpreted to have grown from dehydration reactions, so it was necessary to include $H_2O$ fluid in the list of phases in all simulations.
Fig. 4. Ca, Fe, Mg and Mn element maps for garnets. Lighter shades correspond to higher element concentrations. Lines correspond to the locations of microprobe line traverses. ‘L’ is the left end of the traverse and ‘R’ is the right end, corresponding to Fig. 3.
Determination of segment boundaries

To simulate garnet zoning profiles, each garnet was divided up into segments. Segment boundaries were located at changes in slope along the Ca profile. Points that display prograde reequilibration, as judged from garnet-biotite thermometry calculations (Fig. 3), were avoided. Each segment is simulated in sequence from the core outward. To perform the calculations, values of the monitor parameters $\Delta X_{\text{gr}}$ and $\Delta M_{\text{Gar}}$ were determined for each interval of growth (Table S3).

Obtaining a good fit

The fit of the simulated profile to the analysed profile was judged by visual comparison. Only two parameters, $X_{\text{gr}}$ and the garnet radius (controlled by $\Delta M_{\text{Gar}}$), are guaranteed to fit. In some cases, good fits to $X_{\text{plm}}$, $X_{\text{py}}$ and $X_{\text{gr}}$ were obtained through the trial and error adjustment of three parameters: the nucleation density, the initial value of $X_{\text{Mn-Chl}}$ and $\Delta M_{\text{Gar}}$ (adjusted for each segment). The initial value of $X_{\text{Mn-Chl}}$ makes little difference to the calculated $P-T$ path, but is important because it controls the size of the Mn reservoir, and therefore the rate at which Rayleigh fractionation causes the garnet Mn concentration to drop towards the rim. In the simulations, only slight changes in the compositions accompanied garnet growth for most minerals (Tables S2 & S4). Adjustments were made to the initial compositions to account for changes in those cases where the changes appeared significant (Table S2). Adjustments to modes were also made in some cases (Table S5).

An important complication to consider when making adjustments to initial modes and initial compositions is that all garnets in this study are partially resorbed because of prograde reactions (and in the THAL samples, also because of minor chlorite alteration). The resorption was accompanied by changes to the modes and compositions of the matrix phases, and in some cases, changes to the mineral assemblage. Because only the early portion of the history corresponding to the growth of the garnet, from the core to the resorbed rim, could be simulated, it is not valid to judge the quality of fit by detailed comparisons between the final simulated mineral compositions and modes to the actual values from the rock. Our experience is that simulations are very sensitive to changes in the monitor parameters, but not very sensitive to small changes in the values input for initial matrix compositions or modes, so it is only necessary to know those values approximately.

Simulations using measured values as initial values sometimes failed because of a phase or phase component being completely consumed prior to the completion of garnet growth. This was the result of an insufficient reservoir of a component needed for garnet growth. When this occurred and how it was dealt with is discussed below for the individual cases.

Thal4e

Petrography and reactions

This sample is schist of Mahogany Peaks from the Albion Range and contains garnet, staurolite, muscovite, paragonite, quartz and kyanite. Small grains (5–10 µm) of graphite occur in the matrix, as sparse inclusions in garnet, and as abundant inclusions in staurolite. Ilmenite occurs in the matrix and as inclusions in garnet and staurolite. Accessory rutile and tourmaline occur sparsely throughout. Chlorite occurs only along the rims of garnet as a secondary alteration product. The garnet is porphyroblastic, up to 8 mm in diameter and corroded. Embayments are commonly filled by staurolite (Fig. 5), suggesting a reaction in which staurolite grew at the expense of garnet. The staurolite is porphyroblastic (2–7 mm) and corroded. Small (20–50 µm) kyanite grains are localized along the margins of corroded staurolite, suggesting a reaction that consumed staurolite and produced kyanite. Muscovite, quartz and paragonite make up the fine-grained (<10 µm) matrix. Minor deformation related to flattening and/or shear strains is apparent (Fig. 5).

The sample underwent a complex reaction history that was determined through the interpretation of textural evidence and AFM plots (Fig. 7). The bulk composition of this sample (Table S1) plots above the garnet-chlorite join on an AFM plot, and is thus a high-alumina variety of pelitic schist (Fig. 7). The bulk composition plots approximately in the same location as chloritoid would plot, although no chloritoid was found in this sample. In the pure KFMASH system, chloritoid would be expected to stabilize at some point; however, minor amounts of Ca and Mn in garnet can displace chloritoid stability to conditions outside the realm of ordinary crustal metamorphism (Spear & Cheney, 1989; Wang & Spear, 1991; Mahar et al., 1998). Apart from the effect of Ca and Mn components on chloritoid stability, the KFMASH system is considered to be a good approximation for the purposes of interpreting petrogenesis. Interpreting from the work of Spear & Cheney (1989) in the KFMASH system using AFM plots and the estimated bulk composition (Table S1), garnet grew inside the assemblage chlorite + staurolite (+ paragonite + quartz + muscovite) mainly from the breakdown of chlorite as the three-phase triangle chlorite + staurolite + garnet swept to the left (Fig. 7). With further progradation, the reaction garnet + chlorite = staurolite + biotite was crossed, partially consuming garnet. As a result of this reaction, many embayments along garnet rims became filled with staurolite (Fig. 5). The reaction ceased when all chlorite was consumed. With further progradation, the biotite + staurolite + garnet triangle migrated to the right to encompass the bulk composition, ultimately eliminating biotite from the assemblage while growing.
additional staurolite and garnet (Fig. 7). At this stage, staurolite should have been idioblastic, as no reaction to this point has consumed it. With further progradation, the kyanite + garnet + staurolite triangle swept to the right to encompass the bulk composition, causing kyanite and garnet to grow at the expense of staurolite (Fig. 7). The reaction is confirmed by regression of measured mineral compositions in the rock (Table S1, using the rim composition for G3):

\[ 0.23 \text{pg} + 0.53 \text{st} + 1.61 \text{qtz} = 0.64 \text{grt} + 0.08 \text{ms} + 4.26 \text{ky} + 0.07 \text{ilm} + \text{H}_2\text{O}. \]

The presence of kyanite grains localized along the margins of corroded staurolite grains is consistent with this interpretation.

The above discussion predicts that garnet growth was followed by partial consumption, then by two further episodes of growth. A hiatus in the garnet zoning profiles is plainly seen in all three of the garnets analysed from this sample. Petrographically, the hiatus is represented by concentrations of graphite inclusions that align with discontinuities in the compositional profiles (Fig. 6). A portion of the rim of one garnet was mapped for Y, and displays an annulus consisting of a spike in the Y content just beyond the discontinuity. Yttrium-rich annuli in garnet have been interpreted to result from consumption followed by regrowth when consumption reaction involves the breakdown of minerals that host significant reservoirs of Y (Pyle & Spear, 1999). In this case, the garnet-consuming reaction also consumed muscovite, which is a major host for Y in pelitic schist (e.g. Grauch, 1989). When muscovite was consumed, Y was released into the area along the consuming garnet rim, and then incorporated into the garnet when the garnet regrew. The garnet-consuming reaction releases \( \text{H}_2\text{O} \), and it may be speculated that the fluid dissolved small amounts of graphite from within the matrix and precipitated it into the area along the garnet rim. When the garnet regrew, the graphite that had been deposited along the corroded rim was occluded, forming the graphite concentrations that mark the discontinuity in the profile.

**Thermobarometry**

Well-calibrated thermometers such as garnet-biotite and garnet-chlorite could not be applied to this sample due to the lack of appropriate mineral assemblages. For the same reason, well-calibrated barometers, such as GASP or MBPG, could not be applied. The only reliable constraint on pressure and temperature is the presence of kyanite and the inferred discontinuous reaction \( \text{garnet + chlorite = staurolite + biotite} \) (Spear & Cheney, 1989), which was crossed during progradation. The garnet growth segments to be modelled grew at temperatures below the discontinuous reaction. Consistent with these constraints, conditions of 500 °C and 5 kbar were assumed for the initiation of garnet growth.

**Simulation of garnet growth zoning**

Three garnet from THAL4E were simulated (designated G1, G2 and G3), all from the same polished section. Of
the three stages of garnet growth, only the first stage of growth, which took place within the assemblage chlorite + staurolite + muscovite + paragonite + quartz, could be simulated numerically to determine $P$–$T$ paths. Although the beginning of the second stage of growth, following the hiatus, could theoretically be simulated, there is no feature that would allow one to interpret where the change from the second stage to the third stage occurs. Because the mineral assemblage changed between the second and third stages of garnet growth, and because it is essential to associate the correct portion of the garnet profile with the correct assemblage in the simulations, neither could be simulated. Thus, garnet from this sample was numerically simulated from the cores to the hiatuses. G3 contains a faint irregularly shaped cluster of graphite inclusions near the centre, which corresponds to a distinct area of increased Ca concentration (Figs 3 & 4). This feature is interpreted as a resorbed older garnet, or possibly a relict of an overgrown phase (e.g. Hirsch et al., 2003), and so was avoided in the modelling.

The profile for G1 was well simulated in a single segment of growth, G2 was well simulated in two segments, and G3 was well simulated in seven segments. All three garnet display a high degree of symmetry. G3 provided the most complete path of any single garnet modelled in this study.

In this rock, the reservoir for Ca needed to make the grossular component of garnet is contained entirely in paragonite as the margarite component. When the model was run using measured values as initial values, the simulations ran out of margarite component before garnet growth was completed. In order to provide sufficient Ca for garnet growth, the initial values of both the paragonite composition and modal abundance were adjusted to higher values than determined for the actual rock (Tables S2 & S5).

**Thal6b and Thal6c**

**Petrography and reactions**

These two samples are schist of Mahogany Peaks from the Albion Range (Fig. 2a). Both contain garnet, staurolite, muscovite, biotite, paragonite, plagioclase and quartz. Small grains (<2 μm) of graphite occur in...
Fig. 7. AFM projections (Thompson, 1957) showing reactions responsible for garnet growth. Bulk compositions are shown as stars. (g) Labelled bulk compositions, composition field for normal pelitic chloritoid (grey shaded area) and precisely plotted mineral compositions. Arrows show migration of three-phase triangles in direction of net dehydration. (a–f) Prograde sequence; (a–c) migration of three-phase triangles responsible for initial garnet growth. In samples THAL4E, THAL6B and THAL6C, prograde garnet rim consumption, growth of clear rims on staurolite, and the complete consumption of chlorite resulted from discontinuous reaction grt + chl = st + bt across (c) and (d). Further progradation resulted in consumption of biotite and the corrosion of staurolite in THAL4E as the bt + grt + st triangle migrated to the right (e, f). In THAL4E, texturally late kyanite and re-growth of garnet occurred due to prograde migration of ky + grt + st triangle (e, f). Biotite analyses plotted on triangles in (g) are shown to demonstrate slight offset from assumed biotite compositions. Assumed biotite compositions compensate for low alkali totals that resulted from using small spot size during analysis, necessary due to small grain size. Bt, biotite; Chl, chlorite; Grt, garnet; Ms, muscovite; Ky, kyanite; Pg, paragonite; Pl, plagioclase; Qtz, quartz; St, staurolite.
the matrix, as sparse inclusions in garnet and as abundant inclusions in staurolite. Ilmenite occurs in the matrix and as inclusions in garnet and staurolite. Chlorite occurs only along the rims of garnets as a secondary alteration product. The garnets are porphyroblastic (up to 8 mm in diameter) and embayed. Embayments are commonly filled by staurolite (Figs 5 & 6), suggesting a reaction in which staurolite grew at the expense of garnet. The staurolite is porphyroblastic and subidioblastic. Fine-grained (<10 μm) muscovite, quartz, plagioclase and paragonite make up the matrix. Minor deformation related to flattening and/or shear strains resulted in micro-scale shears of low displacement (Fig. 5).

These samples are similar to THAL4E in that they are a high-alumina variety of pelitic schist, but differ in that the bulk composition is significantly more magnesium-rich (Fig. 7). Using the estimated bulk composition (Table S1) and interpreting the AFM plots of Spear & Cheney (1989), garnet growth is predicted to have occurred via the same reaction as THAL4E, from the breakdown of chlorite as the three-phase triangle chlorite + staurolite + garnet swept to the right (Fig. 7). Garnet growth ended when the reaction garnet + chlorite = staurolite + biotite was crossed, partially consuming garnet. As in THAL4E, the resultant embayments along the garnet rims are commonly filled by staurolite. With further progradation, migration of the biotite + staurolite + garnet triangle to the right partially consumed biotite and grew additional staurolite and garnet; however, unlike THAL4E, biotite is preserved. The additional growth of staurolite is reflected as clear narrow rims on many staurolite grains, overgrowing the inclusion-rich core. The preservation of biotite in these samples is due to the more Mg-rich bulk composition and also to the discontinuous reaction being displaced to higher temperatures than in THAL4E due to the higher concentrations of Ca plus Mn in the garnet cores (Table S1). This interpretation requires that garnet growth began later in these samples than in THAL4E, which is useful information for linking up P–T path segments derived from the different samples (discussed later on). Moreover, unlike THAL4E, staurolite grains are not corroded, because within this bulk composition the prograde staurolite breakdown reaction that affected THAL4E did not occur (Fig. 5).

**Thermobarometry**

Although samples THAL6B and THAL6C contain both garnet and biotite, the biotite grew during a garnet-consuming reaction, and so is not in equilibrium with exposed garnet rims. Consequently, garnet-biotite thermometry calculations would be invalid and so were not made. Because the outcrop for this sample is nearby the outcrop of THAL4E (Fig. 2a), it is reasonable to assume a shared history, and that the constraints available to THAL4E would also apply to these samples. The assumed starting conditions for garnet growth (475 °C for THAL6B-G1 and 490 °C for THAL6B-G2 and 5 kbar for both) are consistent with those constraints.

**Simulation of garnet growth zoning**

Garnet growth was simulated within the assemblage chlorite + staurolite + muscovite + paragonite + plagioclase + quartz. For garnet 3, the initial garnet composition was taken to be the average composition at the margins of the high Ca-feature (shaded area in Fig. 4) and a non-zero starting radius was assumed. THAL6C-G1 was well simulated with a single step (Fig. 3). For THAL6B-G1, only ~80% of the radius of the garnet profile was successfully simulated (Fig. 3). The profile is notably asymmetric, so only the left half was simulated, although corresponding points along both halves were averaged to determine compositions at the segment boundaries. The first 80% of the radius was simulated in two segments of growth, the first yielding a good fit and the second fitting less well, particularly with respect to Fe. Many attempts to simulate the remainder of the profile failed due to the tendency of Fe to drop towards the rim, unlike the actual profile. Although the reason for the inability to simulate this portion of the profile could not be uniquely determined, one possible explanation is open system behaviour with respect to Fe, leading to a bulk increase in the amount of Fe in the rock corresponding in time to the growth of the last 20% of the garnet radius.

In these samples, the reservoir for Ca needed to make the grossular component of garnet is contained in both paragonite as the margarite component and in plagioclase. The garnet growth reaction involved producing paragonite at the expense of plagioclase. In order to prevent the simulation from running out of paragonite, it was necessary to adjust the initial plagioclase mode to a higher value than estimated for the rock (Table S5).

**UH-3 and UH-5**

**Petrography and reactions**

These two samples are pieces of one 20 cm boulder of schist of Stevens Spring from the Basin Creek area in the northern Grouse Creek Mountains (Fig. 2b). They are from the upper horizon of the schist of Stevens Spring, as described in Hoisch et al. (2002). The samples contain the mineral assemblage quartz + muscovite + biotite + plagioclase + garnet, with accessory ilmenite, allanite (mainly as inclusions in garnet), monazite (only in the matrix), apatite and zircon. Ilmenite occurs in the matrix and as inclusions in garnet. There is no chlorite in these samples, neither primary nor secondary. Garnet is porphyroblastic and up to 6 mm in size. Matrix silicate phases are generally
0.2–0.8 mm. Garnets commonly possesses spiral inclusion trails defined by quartz and ilmenite and are corroded along rims (Fig. 5). The corroded rims truncate the interior zoning patterns, except for features related to post-consumption prograde diffusive reequilibration, which caused Fe-enrichment along rims, cracks and some inclusion margins (Fig. 4). The consumed portions of the garnet rims are notably devoid of muscovite, suggesting a muscovite-consuming reaction. Unlike the THAL samples, these contain no paragonite and the matrix is dominated by plagioclase. There is also less late shear deformation, even though the rocks occur within a well-developed Tertiary shear zone. The resistance to shearing may be attributable to the abundance of plagioclase in the sample, which may have provided greater resistance to deformation overprints relative to more mica and quartz-rich adjacent lithologies (e.g. Passchier & Trouw, 2005).

The estimated bulk composition plots below the garnet-chlorite join on an AFM plot (Table S1), indicating a low-alumina (common) variety of pelitic schist. Using the estimated bulk composition and interpreting from the AFM plots of Spear & Cheney (1989), garnet growth is predicted to have occurred via essentially the same reaction as the THAL samples, that is, by the breakdown of chlorite. Garnet corrosion is interpreted to have occurred via a fluid-absent reaction resulting from a major drop in pressure during progradation.

**Thermobarometry**

These samples contain mineral assemblages appropriate for the application of well-calibrated thermobarometry methods, specifically, garnet-biotite thermometry and MBPG barometry. Unlike the THAL samples, garnet is interpreted to have grown in the presence of biotite. For these samples, garnet-biotite thermometry (Hodges & Spear, 1982; Holdaway, 2000) and MBPG barometry (Hoisch, 1991) were calculated using garnet compositions as described below and average matrix compositions for biotite and plagioclase, in an effort to determine the \( P-T \) conditions of initiation of garnet growth (Fig. 8).

The results of the calculations must be interpreted bearing in mind a number of complicating factors. Garnet consumption causes the Fe-content of matrix biotite to increase, causing the calculated temperature to artificially increase (Kohn & Spear, 2000). Garnet consumption also releases Ca from the consumed garnet rim, which becomes incorporated into plagioclase, thus increasing the anorthite content beyond what it was at the end of garnet growth. Because matrix compositions change during garnet growth, combining matrix compositions with garnet core compositions introduces error in thermobarometry calculations. For rocks from this outcrop, Hoisch et al. (2002) estimated that changes in biotite composition accompanying garnet growth could result in temperatures being underestimated by no more than 18 °C. However, associating garnet cores with matrix plagioclases in barometry calculations did not produce good results in this study, so barometry calculations were made using garnet rims (Fig. 8).

Considering these complicating factors, the results should be considered to possess large uncertainties, but garnet rims yielded generally consistent temperatures of 450–460 °C and consistent pressures for the intersections of the barometry equilibria of 4.5–5.5 kbar (Fig. 8). The determined conditions are appropriate for the growth of garnet via chlorite breakdown in pelitic rocks (Spear & Cheney, 1989). For the simulations, the initial pressure was assumed to be 5 kbar and the initial temperature of garnet growth was taken to be the calculated temperatures using the garnet cores at 5 kbar and average matrix biotite assuming 20% of Fe to be \( Fe^{3+} \) (Hoisch et al., 2002).

**Simulation of garnet growth zoning**

Garnet growth in these samples was simulated within the assemblage quartz + muscovite + biotite + chlorite + plagioclase. Two garnets from each of the two samples were analysed and simulated. Garnet-biotite thermometry, plotted for each point along each traverse in Fig. 3, was used as a guide for evaluating whether analysed points were acceptable options for segment boundaries. Points near rims, along cracks, and along inclusion margins display prograde reequilibration, and so could not be used as segment boundaries. Two of the garnets possess profiles that are slightly asymmetric (UH-3-G1 and UH-3-G2), so one half of each garnet profile was chosen for simulation. The profiles of the other two garnets (UH-5-G1 & UH-5-G2) are highly symmetric, and the half with the most complete record was chosen for simulation. UH-5-G2 and UH-3-G1 were each simulated in two segments of growth, while UH-3-G2 required three segments and UH-5-G1 required four segments.

**CREATING THE COMPOSITE \( P-T \) PATH**

In this study, the simulated portions of all garnets grew from very similar reactions involving the breakdown of chlorite. However, because of differences in the bulk chemistry of the rocks and possibly other factors, each garnet initiated growth at a different point along the \( P-T \) path. Because the initial \( P-T \) conditions assumed for the initiation of the growth of each garnet could only be approximated, the generated paths are unlikely to be in their correct relative positions in \( P-T \) space. However, the determined \( P-T \) paths may be shifted slightly in \( P-T \) space without introducing significant errors because the path calculations are not sensitive to small changes in the assumed initial pressure and temperature.
Several approaches were used to determine the relative positions of the \(P-T\) paths in \(P-T\) space. Garnets analysed within the same polished section grew in close proximity to each other, and therefore drew from the same Mn reservoir during growth. This allows Mn concentration along garnets profiles to be used to temporally correlate \(P-T\) path segments from different garnets (e.g. Spear & Daniel, 2003). The correlation of the paths of the three garnets analysed in THAL4E is shown in Fig. 9. The correlation is consistent with garnets growth initiating at different times along the same \(P-T\) path, and with growth of the garnets overlapping to varying degrees. The approach appears to work with the exception of the first segment of the path generated for G2, which is slightly out of place with respect to Mn composition compared with the similarly-sloping final segment of G3. This could be due to the development of slight heterogeneities in the Mn reservoir during garnets growth, such as might occur if subtle Mn depletion halos developed. Similarly, paths may be correlated based on Mn content for the two garnets analysed from UH-3 and the two garnets analysed from UH-5. Each pair of garnets substantially overlap in their Mn contents, and therefore, so should their \(P-T\) paths.

A second approach involves correlating path segments with similar slopes. The repositioning of the three paths from THAL4E based on Mn content described above places into alignment similarly sloping segments (Fig. 9). The same is true of the two garnets from UH-3 and the two garnets from UH-5 (Fig. 10). In addition, the slope of the first segment of the \(P-T\) path from THAL4E-G3 (schist of Mahogany peaks from the Albion range) is in good agreement with similar segments of \(P-T\) paths from the upper horizon of the schist of Stevens Spring from the northern

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**Fig. 8.** Geothermobarometry calculated from analysed mineral compositions. MBPG geobarometry (Hoisch, 1991) and garnet-biotite geothermometry were calculated using garnet rims as follows (see Fig. 3): UH-3-G1 and G2 used point 3, UH-5-G1 used point 4, UH-5-G2 used point 2. Geothermometry (Holdaway, 2000) shown for calculations assuming 10% and 20% of total Fe in biotite is \(\text{Fe}^{3+}\). Geothermometry labelled H&S (Hodges & Spear, 1982) assumes all Fe in biotite is \(\text{Fe}^{2+}\).
Grouse Creek Mountains (Hoisch et al., 2002; and the two UH-5 garnets in this study). The paths of the two garnets from UH-3 were found to be similar to those from the same outcrop reported in Hoisch et al. (2002), in that they possess a similar characteristic inflection where temperatures change from increasing to decreasing. To align them with those in Hoisch et al. (2002), the paths in the current study were shifted as indicated in Table S6. When the starting pressure for the $P$–$T$ paths for UH-5-G1 and G2 is shifted down from 5 to 4 kbar, the paths align with the slope and location of the first path segments of the UH-3 paths, and thus the UH-5 paths appear to represent a cooler, earlier portion of the $P$–$T$ history.

A third approach involves considerations of mineral equilibria. Qualitatively, the interpretation of reactions based on AFM plots indicates that garnet growth in THAL6B and THAL6C should have begun after THAL4E (Fig. 7). In addition, the repositioning of the four UH paths, described above, is supported by the calculated rim temperatures for all four garnets, as all the four paths appear to have ended at similar temperatures (Fig. 8). Conceivably, one could use the Gibbs method of Spear & Selverstone (1983) to determine the relative displacement of paths from rocks that possess the same mineral sub-assemblage representing a thermodynamic variance of three or less, which would allow the calculation to be run on the basis of the differences in garnet core compositions. However, this was not case for the samples in this study.

Using all these approaches, the individual paths were combined into a single composite path in Fig. 10 by shifting each calculated path as specified in Table S6. The distance between samples sites in the northern Grouse Creek Mountains (UH samples, schist of Stevens Spring) and sample sites in the Albion Mountains (THAL samples, schist of Mahogany Peaks) is ~40 km (Fig. 1). The distance between these localities and the uncertainties in the determined pressures of metamorphism allows for significant differences of structural level. Although similar $P$–$T$ path segments for the two rock units most probably record the same tectonic events, the absolute pressures may be somewhat offset from the depiction in Fig. 10. However, major structures that would vertically separate rocks between the two areas have not been identified. Synorogenic low-angle normal faults have been identified in the region (Wells et al., 1998), but the areas of interest lie in the footwalls of these faults and so could not be offset by them.

**TECTONIC IMPLICATIONS OF THE COMPOSITE $P$–$T$ PATH**

The composite $P$–$T$ path overlaps and extends the previously published $P$–$T$ paths from Basin Creek reported in Hoisch et al. (2002). Some general statements regarding tentative correlations between path segments and geological structures and the tectonic significance of these paths can be made, although correlations with mapped geological structures and

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**Fig. 9.** Correlation of $X_{Sp}$ values among garnet from THAL4E. (a) Pressure–temperature plot of $P$–$T$ paths with $X_{Sp}$ values labelled for selected segment boundaries. (b) $X_{Sp}$ values at segment boundaries $v$ inferred time.
deformation fabrics in the region (e.g. Wells, 1997) and with dated events in the fold-thrust belt to the east remain speculative.

Emplacement of the hangingwall rocks of the Basin-Elba fault most probably caused the burial events recorded in the $P$–$T$ paths as well as widespread upper greenschist to amphibolite facies metamorphism of the footwall across the exposure of the Raft River Mountains sequence (M1 in Table 1). The relationship between the thrust at the base of the Robinson Creek sequence and the Basin-Elba fault may indicate that the Basin-Elba fault represents the floor thrust to a duplex fault zone of tectonically thickened Late Cambrian and Neoproterozoic rocks, or that the Robinson Creek sequence represents a faulted upright limb to a fold nappe, with the Mount Harrison sequence representing the lower overturned limb (Miller, 1980, 1983). The Basin-Elba fault accomplished significant crustal thickening as the hangingwall sheet had an estimated stratigraphic thickness greater than 12 km.

By inference, initial motion on the Basin-Elba fault is interpreted to be older than the 105 ($\pm 12$) Ma orogen-parallel extensional fabric, as this fabric is prograde and the metamorphic conditions require significant tectonic burial (Table 1). The Willard thrust sheet, and other ‘megathrust sheets’ that carry thick Proterozoic–Lower Cambrian clastic successions at their bases, including the Canyon Range and Sheep Rock, were initially translated in Early Cretaceous time (144–110 Ma) (DeCelles, 2004). As the Basin-Elba fault lies in a more western position within the orogenic wedge than the Willard thrust (Camilleri et al., 1997), and because it is unlikely that the Basin-Elba fault fed slip into the Willard thrust (see below), we infer that its age of initial motion is older than initial motion on the Willard thrust.

It is unlikely that the Willard thrust, and other thrusts of the foreland fold-thrust belt, received slip from the Basin-Elba fault. The metamorphic rocks in the footwall of the Basin-Elba fault include upper greenschist-lower amphibolite facies Ordovician strata. The Ordovician rocks lie in the hangingwall of the Mahogany Peaks fault, a west-directed low-angle normal fault that post-dates initial motion on the Basin-Elba fault and juxtaposes a hangingwall flat in Ordovician rocks against a footwall flat in Neoproterozoic rocks. The simplest interpretation of these relationships is that the Ordovician metasedimentary rocks were derived from an up-dip segment of the west-tilted footwall of the Basin-Elba fault and displaced down-dip, minimally cutting out the Cambrian section, to rest on the Neoproterozoic schist of Mahogany Peaks. The foreland thrusts sole into a basal decollement in Late Proterozoic siliciclastic rocks to the west, and are interpreted to eventually cut down westward into crystalline basement (Allmendinger, 1992; Yonkee, 1992; Camilleri et al., 1997). Thus, it is unlikely that the basal decollement to the foreland fold-thrust belt buried rocks as young as Ordovician in the study area, requiring a western thrust (e.g. the Basin-Elba fault), now internal to the orogenic wedge, to have buried the metamorphic rocks of the Raft River sequence. The Basin-Elba fault may have fed slip into the Manning Canyon detachment, a hinterland thrust, thought to be Late Jurassic, which forms a bedding-parallel fault over a broad region of northwestern Utah and places Pennsylvanian over Mississippian strata (Allmendinger & Jordan, 1981).

Field geological relationships provide evidence for at least one episode of renewed thrust motion on the Basin-Elba fault following an episode of extension. The extensional Mahogany Peaks fault is folded into an east-vergent, tight to isoclinal, overturned syncline in the immediate footwall of the Basin-Elba fault (Miller, 1980, 1983), suggesting a kinematic link between east-vergent folding and motion on the fault (D3 in Table 1). We tentatively correlate the youngest segment of the composite $P$–$T$ path, a pressure increase, with this renewed thrusting event. The age constraints of renewed thrusting, $c$. 60–45 Ma, deter-

**Fig. 10.** $P$–$T$ paths generated from garnet growth simulations and shifted as indicated in Table S6. $Al_2SiO_3$ stability fields from Pattison (1992); And, andalusite; Ky, kyanite; Sil, sillimanite. (a) Paths from schist of Stevens Spring samples in Basin Creek. (b) Paths from schist of Mahogany Peaks samples, Albion Mountains. (c) Paths from all samples. Tentative correlations of path segments to tectonic events (Table 1) are shown.
mined by Th-Pb dating of monazite inclusions in garnet (revised from Hoisch and Wells 2004), are consistent with garnet growth following motion on the Mahogany Peaks fault that is bracketed between c. 90 and 60 Ma (Table 1).

The exhumation events recorded in the composite P–T path may have occurred by a combination of extensional reactivation of the Basin-Elba fault (Hodges & Walker, 1992), motion along the extensional Mahogany Peaks and Emigrant Spring faults (Wells, 1997; Wells et al., 1998), extensional flow associated with mid-Cretaceous orogen-parallel extension or along structures not yet recognized. We tentatively correlate the two major decompressional segments of the path with motion along the Mahogany Peaks and Emigrant Spring extensional faults, because the Proterozoic pelitic schists lie in the footwall of both and this is permitted by the available geochronology. The temperature increase of ~50 °C that accompanied the first decompression is attributable to thermal relaxation associated with thrusting along the Basin-Elba fault (Hoisch et al., 2002); however, heating following the second decompression may require an additional heat source, such as might result from lower lithospheric delamination.

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**SUPPLEMENTARY MATERIAL**

The following supplementary material is available for this article online http://www.blackwell-synergy.com:

**Table S1.** Mineral and calculated bulk compositions.

**Table S2.** Parameters used in garnet growth simulations.

**Table S3.** Monitor parameters, $\Delta X_{gr}$ and $\Delta M_{gar}$, used in garnet growth simulations.

**Table S4.** Measured $v$ simulated mineral compositions.

**Table S5.** Initial and final mineral modes from garnet growth simulations, compared with visually estimated modes.

**Table S6.** $P$–$T$ path displacements.