The history of anatectic pelites of the Northern East Humboldt Range, Nevada: Evidence for tectonic loading, decompression, and anatexis

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The migmatitic lower plate of the Ruby Mountains–East Humboldt Range (RM–EHR) metamorphic core complex of northeastern Nevada represents a deep section of the hinterland of the Sevier orogenic belt. New major and trace element (Zr-in-rutile, Ti-in-biotite) thermobarometry is aided by garnet zoning analysis and a thermodynamic modeling approach involving melt reintegration. The P–T history for rocks from the Winchell Lake nappe (WLN) is characterized by a phase of high-temperature, nearly isothermal, probably tectonic loading followed by decompression and melting during continued heating from 675°C to 740°C. Rocks of the Lizzies Basin Block (LBB) below the emplacement fault for the WLN allochthon record no evidence of a nearly isothermal phase of tectonic loading and associated metamorphism. Hence a different P–T path characterized by more widespread melting during prograde heating and contraction is compiled for the LBB. Different interpretations are presented for the exhumation history and the context of partial melting in the WLN versus the LBB. Anatexis in pelites of the WLN occurred as a result of dehydration melting reactions that took place in a decompressional phase of the prograde P–T path, in contrast to dehydration melting during contraction in the LBB. The emplacement of the WLN must have been a significant event in the tectonic history of the RM–EHR. These results have important implications regarding the relationship between crustal thickening, syncontractional exhumation, and anatexis within continental orogenic systems.

Key words: migmatite; garnet; partial melting; metamorphic core complex; zirconium; rutile

INTRODUCTION

In the lower to middle crust, partial melting of metasedimentary rocks dominantly occurs as a result of dehydration melting reactions of hydrous phases, most commonly micas, in conjunction with quartz and plagioclase. Anatexis and migmatization require some combination of heating and/or decompression across the effective solidus (e.g. Groppo et al., 2012), which has a steeply positive slope in pressure–temperature (P–T) space. The slope of the P–T path crossing this solidus is critical to the interpretation of the tectonic evolution that led to anatexis, as well as the tectonic implications of the production of significant volumes of partial melt.

The presence of even a small volume of distributed partial melt (several per cent) can have a profound effect on the bulk rheological properties of a crustal section of rock (Arzi, 1978; Paquet et al., 1981; Hollister & Crawford, 1986; Rushmer, 1996; Rosenberg & Handy, 2005; Vanderhaeghe, 2009). Anatexis and leucogranite generation are demonstrated to play a significant role in crustal differentiation (e.g. Brown, 1994), in addition to contributing to a decrease in bulk-rock strength. Studies in the Himalaya have pointed to anatexis and leucogranite magmatism as both a trigger (e.g. Searle et al., 2010) and a result of exhumation of a mid-crustal section or ‘channel’ (Patiño Douce & Harris, 1998; Harris et al., 2004). Decompression of hot crust was, at least in part, a result of syncontractional normal-sense motion on the South Tibet Detachment.
system (Harris & Massey, 1994; Patiño Douce & Harris, 1998; Harris et al., 2004; Searle et al., 2010).

Rocks of the hinterland of the Sevier Orogenic belt in western North America record tectonism related to crustal thickening and contraction along an Andean-style continental margin. The nexus of crustal thickening occurred west of the preserved Sevier fold and thrust belt at the western edge of the North American craton, in a region that has since experienced protracted Cenozoic magmatism and extension. Exposures of the Sevier hinterland are limited to exhumed footwall blocks in a series of metamorphic core complexes stretching from Mexico to the Canadian Cordillera (Fig. 1; Armstrong, 1982). Modern surface expressions of this mid-crustal belt are much more limited south of the Snake River Plain. Miocene to present Basin and Range extension has exposed the footwall block of the Ruby Mountains–East Humboldt Range (RM–EHR) metamorphic core complex in northeastern Nevada. The RM–EHR represents a section of the middle to lower crust affected by Sevier tectonism and associated magmatism (Fig. 2). The $P$–$T$ and melting history of these rocks has significant bearing on the interplay of deformation and anatexis as they relate to the evolution of an orogenic system.

This study presents new petrological data as well as major and accessory phase thermobarometric analysis to improve constraints on $P$–$T$ paths for migmatites of the northern East Humboldt Range. In addition, the prograde and retrograde reaction history based on complex major and trace element zoning in garnet and thermodynamic modeling is presented. This contribution addresses the cause and significance of partial melting as it relates to decompression and heating. The results have implications regarding the nature of thickening and the mechanisms of exhumation of continental crust with parallels to active contractional orogenic systems such as the Himalaya and Andes.

**GEOLOGICAL SETTING**

The Ruby Mountains–East Humboldt Range metamorphic core complex (Davis & Coney, 1979), like other Cordilleran core complexes, is characterized by unmetamorphosed supracrustal rocks separated from metamorphic and intrusive rocks by a low-angle, top-to-the-NNW, mylonitic ductile shear zone and associated brittle normal fault system. The infrastructure, or footwall of the core complex shear zone, contains abundant leucogranitic dikes and sills among other Jurassic to Tertiary intrusions. The leucogranite dikes and sills are concordant with, and cross cut the preserved subhorizontal metamorphic fabric. At least one generation of leucogranites has been interpreted to be genetically linked with distinct in situ leucosomes in gneiss and metapelite, suggesting that the magmatic and metamorphic history of this range are inherently linked (Batun, 1999; McGrew et al., 2000; Lee et al., 2003).

Protoliths of the RM–EHR metamorphic infrastructure include Archean–Paleoproterozoic basement gneisses and Neoproterozoic–Paleozoic miogeosynclinal sedimentary rocks including siliciclastic, calcareous, and pelitic rocks (Howard, 1980; Lush et al., 1988; McGrew et al., 2000). Two Jurassic granitoid plutons are preserved in the Ruby Mountains (Hudec, 1992; Jones, 1999b; Howard et al., 2011), yet no similar intrusions are recognized in the East Humboldt Range. A suite of granitoid intrusive rocks of Cretaceous to Early Oligocene age is ubiquitous throughout the RM–EHR infrastructure. In the Ruby Mountains, these rocks include possibly five generations of leucogranites (c. 92 Ma, possibly 83 Ma, 69 Ma, 38 Ma, and 29 Ma), an Eocene biotite–hornblende–quartz diorite, and Oligocene biotite monzogranites ($U$–$Pb$ zircon and monazite ages; Wright & Snoke, 1993; MacCready et al., 1997; Howard et al., 2011). All of these intrusions are penetratively deformed by top-to-the-NNW mylonitic shearing.

**Fig. 1.** Sketch map of part of the western North American Cordillera showing the Ruby Mountains–East Humboldt Range (RM–EHR) metamorphic core complex (black box), one of few exposures of the root of the Sevier orogenic belt in the Great Basin; ARG, Albion–Raft River–Grouse Creek complex; SR, Snake Range complex (after Davis & Coney, 1979; Foster et al., 2007).
associated with the main core complex shear zone, suggesting that plutonism preceded or was synchronous with core complex extension.

Contractional deformation is widespread in RM–EHR infrastructure, manifested as large-scale, recumbent, isoclinal folds (i.e. fold nappes) and a penetrative axial planar foliation (Snoke et al., 1979; Howard, 1980; Lush et al., 1988; Snoke & Miller, 1988; MacCready et al., 1997; McGrew et al., 2000). Upper amphibolite- to granulite-facies metamorphism, and possibly coeval intrusion of
pegmatitic leucogranite, has been interpreted as syn-kine-
matic with respect to emplacement of the Winchell Lake
fold nappe (WLN) in the East Humboldt Range (Hodges
et al., 1992; McGrew et al., 2000; Fig. 3). Late Cretaceous
U–Pb ages for zircon rims in the migmatitic Angel Lake
orthogneiss from the core of the WLN have been inter-
pred to suggest that migmatization and melt crystalliza-
tion occurred at 91 ± 72 Ma (Premo et al., 2008, 2010;
McGrew & Snoke, 2010). The high-grade phase of tecton-
ism is obscured by a lower grade overprint of biotite +
sillimanite-grade extensional tectonism, evidenced by
localized fibrolitic sillimanite and biotite intergrowths
with dynamically recrystallized quartz in extensional
shear bands (McGrew et al., 2000). Most geothermometers
were partially reset on the retrograde path, and, as a
result, peak temperature and temperatures at which peak
pressure was attained are difficult values to constrain.

A possible phase of Late Cretaceous decompression in
the significantly thickened Sevier hinterland has been
discussed in previous studies of Sevier-related metamorphic
belts (e.g. Hodges & Walker, 1992; Camilleri & Chamber-
lain, 1997; Hoisch et al., 2002). In the RM–EHR, some
workers have suggested that Late Cretaceous decompres-
sion is due to a phase of syncontractional ‘tectonic denuda-
tion’ (Hodges et al., 1992), perhaps representative of
gravitational collapse of overthickened crust (Hodges
et al., 1992; McGrew & Snee, 1994; Camilleri & Chamber-
lain, 1997; McGrew et al., 2000). Surface expressions of
Late Cretaceous extensional structures occur in several
parts of the Sevier hinterland (Camilleri & Chamberlain,
1997; Druschke et al., 2009a, 2009b). Parts of the Sevier
hinterland contain sedimentary basins with moderately
thin Late Cretaceous to Eocene deposits (Vandervoort
& Schmitt, 1990) and buried normal faults with relatively
low-magnitude offset (Long, 2012). These are arguably
minor features and have been suggested by some workers
to be consistent with only a small degree of Late Cret-
aceous to Eocene upper crustal extension (Long, 2012;
Miller et al., 2012). The later, core complex phase of exhum-
ation occurred primarily in the Oligocene as mylonitic
extensional shearing of the main core complex shear
zone (Hurlow et al., 1991; Hodges et al., 1992). The full
exhumation history of the high-grade metamorphic infra-
structure of the RM–EHR therefore includes two major
phases, (1) Late Cretaceous and (2) primarily Oligocene
core complex formation.

The relationship between plutonism, contractional
deformation, and decompression–exhumation is complex.
Episodic Late Cretaceous to Oligocene magmatism
(Wright & Snoke, 1993; Premo et al., 2000; Howard et al.,
2011) overlaps 40Ar/39Ar thermochronological constraints
for exhumation related to extensional core complex forma-
tion in the Late Eocene to Miocene (Dallmeyer et al.,
1986; Dokka et al., 1986; McGrew & Snee, 1994). Cross-
cutting intrusive relationships with plutonic rocks both cut-
ting and affected by penetrative contractional deformation
are well documented in the RM–EHR (MacGready et al.,
1997; McGrew et al., 2000; Howard et al., 2011). These field
relationships suggest that plutonism probably began in a
contractional regime, prior to the Late Cretaceous phase
of exhumation, and continued, episodically, through the
initiation of the core complex phase of extension.

PREVIOUS PETROLOGICAL
STUDIES
Various thermobarometric methods have been applied in
efforts to constrain a P–T path for the RM–EHR and adja-
cent ranges. Hodges et al. (1992) applied the Gibbs
method (Spear & Selverstone, 1983) using biotite and
plagioclase inclusions in zoned garnet porphyroblasts to
model P–T paths for rocks from the northern Ruby
Mountains, southern East Humboldt Range, and the
Clover and Wood Hills areas (see Fig. 2). They proposed
distinctly different P–T paths for different tectonic units:
rocks adjacent to the RM–EHR core complex mylonite
zone reached ~750°C at ~6 kbar, whereas kyanite-bearing
rocks from the Clover and Wood Hills area are dominated
by high-pressure metamorphism and steep, nearly isother-
nal decompression from pressures near ~10 kbar at
~600°C (Hodges et al., 1992). P–T conditions during exten-
sional mylonitization were constrained by assuming
garnet porphyroclast rims were in equilibrium with recryst-
ialized matrix biotite, plagioclase, and muscovite, yield-
ing ~600°C and ~3.5 kbar (Hurlow et al., 1991). Estimates
of peak conditions from the Wood Hills area were inferred
from the T–XCO2 relationships of calc-silicate assemlages,
suggesting highest grade peak temperatures of 620–640°C
at ~6 kbar in diopside + dolomite + quartz-bearing rocks
(Camilleri & Chamberlain, 1997). The Clover Hill area
(see Fig. 2) is generally considered a block that was con-
tinuous with, but fault separated from, the EHR. Peak
metamorphic conditions at Clover Hill were estimated at
~7.2 kbar and ~720–760°C (Sicard et al., 2011), though
abundant kyanite implies that prograde metamorphism
passed through the kyanite field below peak conditions,
consistent with a higher pressure phase of metamorphism
(Hodges et al., 1992; Snoke, 1992).

Thermobarometry and stable isotope geochemistry were
performed on rocks from the Lizzlies–Wecks Basin area
(Fig 3) in the northern East Humboldt Range (Wickham
focused mainly on marble and calc-silicate units. Peak con-
ditions were estimated at ~6 kbar and 600–750°C, with a
retrograde overprint that resulted in amphibole + epi-
dote ± garnet assemblages replacing clinopyroxene and
plagioclase in calc-silicate gneisses (Peters & Wickham,
1994). Thermobarometric determinations from metapelite
and metabasic rocks from the WLN and the LBB yield peak conditions of \( \approx 9 \) kbar and \( \approx 800 \) °C (McGrew et al., 2000). An array of internally consistent thermobarometry results is interpreted to define the retrograde portion of the \( P^T \) path characterized by decompression from peak conditions for the collective northern portion of the East Humboldt Range (McGrew et al., 2000).

Geochemical studies of intrusive rocks in the RM–EHR indicate that the Late Cretaceous peraluminous leucogranites have compositions that are enriched in \( \delta^{18}O \) (whole-rock \( \delta^{18}O \) \( \approx 12\%o \); Hodges et al., 1992; McGrew et al., 2000) and depleted in heavy rare earth elements (HREE) with little or no Eu anomaly (Batum, 1999; Lee et al., 2003). Petrogenetic modeling of these rocks suggests that they are derived from partial melting of metasedimentary source rocks. Lee et al. (2003) modeled a pegmatitic leucogranite prevalent in the Lamoille Canyon area of the Ruby Mountains (Fig. 2), interpreted to be genetically similar to leucogranites in Lizzies–Weeks Basin (Fig. 3; Batum, 1999). The models suggest that these leucogranites formed primarily as a result of muscovite dehydration melting of metapelitic rocks, possibly the Neoproterozoic McCoy Creek group mapped in the area (Hudec, 1990; Jones, 1999a). This is consistent with evidence for fluid-absent anatexis of micaceous metapelites discussed below.

![Figure 3](image-url)
METAPELITES OF THE NORTHERN EAST HUMBOLDT RANGE

Metapelitic schists, paragneisses, and migmatitic rocks were collected from within the recumbent south-verging Winchell Lake nappe (WLN) and structurally beneath the nappe in the Lizzies Basin block (LBB). Samples EH06–EH10 are from within a graphitic schist and quartzite unit mapped by A. J. McGrew (McGrew, 1992; McGrew et al., 2000), interpreted as the youngest pre-metamorphic metasedimentary unit exposed in the East Humboldt Range. These samples are locally migmatitic (Fig. 4a) and contain pegmatitic leucogranite intrusions (Fig. 4b), with only EH09 and EH10 showing distinct leucosomes that are overprinted by penetrative shear fabrics and a lower amphibolite-facies retrograde plagioclase + biotite + sillimanite assemblage. Sample EH21 is a garnet-bearing mylonitic paragneiss with a restitic composition from lower in the stratigraphic section. The protolith of this rock is probably of Proterozoic age (Angel Lake Paragenesis of McGrew et al., 2000). Quartz ribbons and prominent shear bands define a penetrative mylonitic foliation. Sample EH22 was collected from the edge of a garnetiferous vein rich in anorthitic plagioclase that is texturally similar to the retrograde calc-silicate assemblages described by Peters & Wickham (1994).

Field mapping (McGrew, 1992; McGrew et al., 2000) of the lower limb of the WLN delineates a series of low-angle fault contacts, which are concordant with the penetrative fabric and probably formed during Late Cretaceous tectonism. These structures are partly obscured by ubiquitous leucogranite intrusions of Late Cretaceous to Tertiary age. For this study, rocks that lie structurally below the mapped fault–shear zone that cuts across the nose of the WLN above Winchell Lake (the ‘main WLN emplacement fault’) are considered to be part of the LBB and rocks above this fault–shear zone are part of the WLN. Samples from the LBB (EH30, EH31, EH37, EH43, EH46, and EH49) were collected from paragneiss units and from within a garnet-biotite-rich, locally migmatitic, schist layer in the Lizzies–Weeks Basin (Fig. 3). This area has been the subject of several petrological studies (Wickham & Peters, 1990; Peters & Wickham, 1994, 1995; Batum, 1999; McGrew et al., 2000). Samples EH43, EH46, and EH49 are from a garnet + biotite schist unit in Weeks Basin that is locally migmatitic. The leucosome domains do not occur strictly as foliation-parallel, sub-planar layers, but rather as irregular, commonly elongate centimeter-scale bodies that are interpreted as amalgamations of partial melt formed in situ (Fig. 4c, d and e). At one outcrop the leucosome material appears to form a web-like network (Fig. 4f).

Samples EH30, EH31, and EH37 from garnet-bearing paragneiss contain both mafic (biotite-rich) and felsic bands. A small number of thin stringers of quartz + plagioclase are interpreted as leucosomes in EH30, suggesting that some in situ melt also formed in the paragneiss. Felsic bands differ from the quartz + plagioclase stringers in that they contain a significant amount of biotite and coarse feldspar porphyroblasts. Biotite within the felsic bands appears to have a more random orientation than in the mafic layers, which may indicate that it was late to grow and formed as a result of melt crystallization reactions.

MINERAL ASSEMBLAGES AND METAMORPHIC TEXTURES

Winchell Lake nappe

Metapelitic samples from the WLN (EH06, EH09, EH10, and EH21) are characterized by a peak assemblage of garnet + biotite + sillimanite + ilmenite + plagioclase + quartz + apatite + monazite + zircon ± tourmaline ± graphite (+ melt). Relict kyanite porphyroblasts are present in samples EH06–EH10 (e.g., Fig. 5a). Staurolite inclusions in garnet are present in samples EH10 and EH21 (Fig. 5b–d). Rutile occurs as inclusions in garnet (Fig. 5c) and kyanite, and as a few coarse grains in the matrix.

Sample EH22 contains kyanite inclusions in grossular-rich garnet (Fig. 5f) along the margin of a calc-silicate layer. This garnet may have formed during retrograde reactions involving leucogranite intrusions that led to the formation of epidote + garnet ± amphibole assemblages in the calc-silicates (Peters & Wickham, 1994). However, no kyanite was found in adjacent metapelitic sample EH21 and is not present in most other parts of the Winchell Lake nappe (McGrew et al., 2000), apart from the graphitic schist unit (samples EH05–EH10). No distinct post-metamorphic structure separates samples EH09 and EH10 from EH21. Sample EH21 is characterized by a strong shear fabric and dynamically recrystallized quartz ribbons, and contains sillimanite and abundant biotite (~50%), suggesting that intense deformation may have helped to transform relict kyanite to sillimanite during enhanced retrograde metamorphism on the exhumation path. Matrix kyanite in the graphitic schist samples (EH06, EH09, EH10) in some places is separated from quartz by a reaction rim of plagioclase (Fig. 6a). Other matrix grains are surrounded by biotite and commonly overgrown by fibrous and prismatic sillimanite (Fig. 6b).

Distinct zones or layers of nearly equigranular quartz and plagioclase (± tourmaline) are present in samples EH09 and EH10 (Figs 4a and 6c, d). These are commonly separated from the melanosome by biotite-rich selvages, and may represent leucosomes formed during partial melting. The rest of the rock has a more Fe- and Mg-rich bulk composition, rich in biotite, garnet, and sillimanite, which probably represent melanosomes and mesosome material surrounding leucosomes. Significant reaction coronas of...
plagioclase + biotite + sillimanite are present around portions of nearly all melanosome garnet in EH09 and EH10 (Fig. 6e). In some places, sillimanite + biotite pseudomorphs after garnet are observed. No muscovite is observed in the graphite schist samples (EH06–EH10) or the mylonitic paragneiss (EH21).

**Lizzies Basin Block**

Locally migmatitic metapelitic samples from the LBB contain the peak metamorphic assemblage garnet + biotite + sillimanite + plagioclase + alkali feldspar + quartz (± melt) + ilmenite + apatite + monazite + zircon. Pelitic schist samples EH45, EH46, and EH49 are richer in biotite and sillimanite than paragneiss samples EH30, EH31, and EH37, which exhibit a gneissic texture. Garnet porphyroblasts in LBB metapelites contain inclusions of the prograde phases biotite, quartz, plagioclase, rutile, apatite, monazite, ilmenite, and rare xenotime. Xenotime inclusions, where present, are limited to garnet cores, although some xenotime grains are found along

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**Fig. 4.** Field photographs and digital slab scans showing leucosome textures. (a) *In situ* leucosome material in graphitic schist on the upper limb of the WLN. (b) Muscovite-bearing pegmatitic leucogranite in WLN graphitic schist. (c) Amalgamation of leucosome (arrow) in garnet + biotite schist of Lizzies–Weeks Basin. (d) Lineation normal cut showing irregular leucosome body (arrow) in hand sample from Lizzies–Weeks Basin. (e) Elongate *in situ* leucosome, sample EH48 from Lizzies–Weeks Basin. (f) Web-like interconnected network of leucosome in garnet + biotite schist, Lizzies–Weeks Basin.
garnet rims, within plagioclase-rich zones that may represent reaction rims. Some garnet porphyroblasts, particularly those adjacent to or within quartzofeldspathic layers that may be leucosomes, contain an inclusion-poor mantle zone surrounding an inclusion-rich core.

Kyanite and staurolite are not present in any of the samples from the LBB. Rutile occurs mainly as inclusions in garnet, but some matrix rutile is present along the edges of ilmenite associated mainly with biotite and muscovite. LBB schist units are dominated by domains rich in biotite and

Fig. 5. Photomicrographs of relict phases within the WLN. Abbreviations after Kretz (1983). (a) Metastable kyanite porphyroblast in sample EH09 matrix. (b) Staurolite inclusion in garnet. (c) Crossed polarized light image showing staurolite and biotite inclusions in garnet. Biotite inclusions in garnet core define an earlier fabric (S\textsubscript{i}), discordant to the mylonitic foliation in the matrix. (d) Ilmenite, quartz, and staurolite inclusions in poikiloblastic garnet. (e) Relatively coarse rutile inclusions in garnet occur along with fine (<10 \mu m thick) rutile needles (not visible at this scale). (f) Ilmenite and kyanite inclusions in garnet porphyroblast.
sillimanite with quartzofeldspathic layers forming ~10–20% by volume. These felsic layers are separated from the more mafic assemblage by biotite- and muscovite-rich selvages in one sample (EH30; Fig. 7a). Muscovite is present both as grains parallel to the dominant foliation and as grains that appear to cross-cut the fabric. The volumetric proportion of muscovite is generally highest within ~10 mm of quartzofeldspathic layers that may have been leucosomes.

Reaction textures in LBB metapelites are generally consistent with retrograde processes. Garnet porphyroblasts in some samples are irregularly shaped and strongly embayed (Fig. 7b and c), surrounded by zones of nearly...
equigranular quartz + plagioclase + biotite. Prismatic sillimanite in some samples shows pull-apart structures filled by muscovite + plagioclase (Fig. 7d). Most alkali feldspar present occurs in leucosomes, where it is commonly associated with muscovite and plagioclase showing myrmekitic intergrowths of quartz (Fig. 7e).

**MINERAL CHEMISTRY AND ZONING IN METAPELITE**

The results of geochemical analysis of key metamorphic phases, including mineral zoning, performed by electron microprobe, are discussed below. Detailed analytical
methods and results are summarized in Electronic Appendix 1 (downloadable from http://www.petrology.oxfordjournals.org).

Garnet

Metamorphic garnet is particularly useful in deciphering the petrogenetic history of complex metamorphic terranes because it is refractory (Hollister, 1969) and grows as a consequence of a variety of metamorphic reactions (e.g. Tracy et al., 1976; Spear, 1995). At metamorphic temperatures, major and trace elements in garnet diffuse at different rates, allowing the mineral to preserve growth zoning in some elements whereas other elements are diffusional modified owing to changes in intensive and extensive variables (e.g. Hickmott et al., 1987; Spear & Florence, 1992; Caddick et al., 2010).

Winchell Lake nappe

Garnet porphyroblasts in the graphitic migmatitic schist samples (EH09 and EH10) are ~2–10 mm in diameter and generally almandine rich ($X_{\text{Alm}} = 0.68–0.75$). They are characterized by variably preserved, diffusional ‘relaxed’ prograde growth zoning that reflects different length scales for diffusion of major and trace elements. This is manifest in most grains as a high-$X_{\text{Sp}}$ (0.04–0.07) and high-$\text{Fe/Fe+Mg}$ (0.85–0.87) core and rimward decrease in these components (Fig. 8a and b). Garnet porphyroblasts with no or poorly preserved higher $X_{\text{Sp}}$ and $\text{Fe/Fe+Mg}$ cores also occur in samples EH09 and EH10, generally restricted to what are interpreted as leuosome domains. One garnet grain in sample EH10 contains a partially preserved relatively high-$X_{\text{Sp}}$ core, but no equivalent high $\text{Fe/Fe+Mg}$ zone. Intra-sample variations in major element zoning magnitudes may in part reflect thin-section cuts that do not sample the center of the garnet porphyroblasts.

Distinct mantles are observed in garnet Ca zoning in EH09 and EH10, showing a partly abrupt increase in $X_{\text{Grs}}$ content toward the rim (from ~0.085 to 0.095), which is more defined in some porphyroblasts (see Figs 8c and 9d). $X_{\text{Grs}}$ then continues to rise to a peak (~0.1) towards the rim, which may reflect garnet growth at the peak pressure experienced by these rocks. In one sample (EH10a-p, garnet l), the garnet mantle is characterized by a discrete step in Ca content that is only partly preserved (Fig. 9b). This feature represents the second significant increase in Ca content and may signal garnet growth during one of two scenarios: (1) a change in the bulk mineralogy of the rock, such as when melting reactions began to break down plagioclase; or (2) a rapid increase in pressure during a phase of garnet growth resulting in a major shift in garnet + plagioclase Ca partitioning. If the higher $X_{\text{Grs}}$ mantles represented garnet growth during anatexis, with Ca supplied to garnet during continuous breakdown of plagioclase, then an abrupt jump in $X_{\text{Grs}}$ and a continued Rayleigh fractionation-like $X_{\text{Grs}}$ peak or annulus would be expected (Spear et al., 1999), whereas lower $X_{\text{Grs}}$ garnet growth would continue beyond this Ca peak. This is not observed, and in most garnets the highest $X_{\text{Grs}}$ zones are smooth. As noted above, significant shifts in the $X_{\text{Sp}}$ and $\text{Fe/Fe+Mg}$ occur within the higher $X_{\text{Grs}}$ mantles. High-$X_{\text{Grs}}$ mantles are also discontinuous, suggesting that significant resorption took place prior to growth of a lower $X_{\text{Grs}}$ rim (see arrows in Fig. 8c). Lower $X_{\text{Grs}}$ (0.05–0.07) rims are cut by resorption textures (Fig. 9b and d). This garnet may be relict ‘anatectic’ garnet that grew during melting.

The near-rim major element zoning pattern in garnet from the graphitic migmatitic schist samples includes troughs of 0.01–0.04 $X_{\text{Sp}}$, and 0.80–0.82 $\text{Fe/Fe+Mg}$. There is then a strong increase in $X_{\text{Sp}}$ and $\text{Fe/Fe+Mg}$ toward the rim, with maximum values that vary between grains, reaching as high as 0.125 and 0.675 respectively. The variation in $X_{\text{Sp}}$ may be due to local kinetic factors during retrograde cooling and garnet resorption, possibly related to variations in leucosome volume during retrogression.

Mylonitic paragneiss EH21 contains two types of garnet porphyroblast. Poikiloblastic garnet shows a simple zoning pattern that is drastically modified by resorption, with a relatively low Ca content (~0.05; Fig. 10a) and sharp increases in $X_{\text{Sp}}$ and $\text{Fe/Fe+Mg}$ at the rim (from ~0.02 to 0.07 and from ~0.87 to 0.92 respectively; Fig. 10b and c). Resorption of poikiloblastic garnets is probably greatly affected by the increased surface area when former inclusions and their grain boundaries are connected to the matrix. Poikiloblastic garnet may represent garnet grown during anatexis.

Sample EH21 also contains large garnet porphyroblasts with fewer inclusions. Despite the fewer inclusions, these garnets contain an important inclusion suite that records the prograde history of these rocks. Staurolite, ilmenite, biotite and rutile (Fig. 5c and d) inclusions indicate that these garnets grew in the presence of staurolite and probably rapidly during the breakdown of staurolite and possibly chlorite. A staurolite inclusion occurs in the garnet core, which is high in $X_{\text{Grs}}$ (~0.095; Fig. 10d) and relatively low in $X_{\text{Sp}}$ (~0.02; Fig. 10e), and $\text{Fe/Fe+Mg}$ (~0.845; Fig. 10f). Porphyroblastic garnet rims show an important and significant decrease in $X_{\text{Grs}}$ to ~0.075 with a coincident increase in $X_{\text{Sp}}$ (~0.06), and $\text{Fe/Fe+Mg}$ (~0.93). This lower $X_{\text{Grs}}$ rim is interpreted to represent garnet grown during anatexis. The $\text{Fe/Fe+Mg}$ increase near the rim is not a function of the rim grain boundary phase (i.e. not higher along garnet + biotite grain boundaries). These patterns are similar to those in garnet from samples EH09 and EH10, although no lower $X_{\text{Grs}}$ cores are preserved in EH21 garnet.

Garnet within the WLN preserves several different zoning patterns owing to two main processes: (1) garnet...
nucleation occurred at several times during the metamorphic history of these rocks; (2) garnets in different textural settings within the same sample preserve different zoning patterns, indicative of the localized scale (less than thin-section scale) of metamorphic equilibration. The difference in the zoning patterns between the graphitic migmatitic schist (EH09 and EH10) and the mylonitic paragneiss (EH21) can be interpreted to indicate that some garnet nucleation in EH21 did not occur until staurolite breakdown at relatively high pressure, and hence EH21 garnet cores grew during the same higher pressure conditions at which the high-$X_{Grs}$ garnet mantles grew in EH09 and EH10.

Dramatic Y zoning is seen in samples EH09 and EH10, characterized by a high-Y core that shows a Y increase to a diffuse annulus and then a slight decrease followed by a sharp decrease that corresponds to the core–mantle increase in $X_{Grs}$ discussed above (Fig. 8c and d). The significance of the Y annulus, which is not observed in other garnets, may be due to (1) kinetic factors affecting the supply of Y to the garnet rim during garnet growth or (2) progressive garnet growth in equilibrium with xenotime during a relatively steep pressure increase. The drop-off in Y from an annulus as high as 4400 $\pm$ 130 ppm (see Fig. 8d) to values below the electron microprobe detection limit (<200 ppm) marks the disappearance of xenotime as a Y ‘buffer’ during garnet core growth (Pyle & Spear, 1999). This assertion is consistent with the presence of xenotime inclusions in the garnet core (see Fig. 8f), but not in the low-Y mantle. An increase in Y along garnet rims (to $\sim$500 ppm) may represent garnet growth during melting reactions that broke down monazite. The higher Y rims correspond to the lower $X_{Grs}$ rims discussed above.
**Lizzies Basin Block**

Garnet porphyroblasts from paragneisses and migmatitic pelitic schists from the LBB are generally almandine rich \(X_{\text{Alm}} = 0.72-0.78\). Increases in \(X_{\text{Sps}}\) and \(\text{Fe}/(\text{Fe} + \text{Mg})\) occur at the rim of all metapelitic garnet from the LBB, suggesting that at least some garnet resorption occurred along the retrograde path. Most grains show cores that are nearly homogeneous in \(X_{\text{Sps}}\) and \(\text{Fe}/(\text{Fe} + \text{Mg})\) (Fig. 11a, b, d and c). Intergrain variation in \(X_{\text{Sps}}\) and \(\text{Fe}/(\text{Fe} + \text{Mg})\) core compositions is consistent with the interpretation that diffusion during both prograde and retrograde metamorphism had a profound affect on the major element zoning. LBB garnet does not preserve the distinct lower \(X_{\text{Grs}}\) cores that are found in WLN garnets.

Garnet porphyroblasts in locally migmatitic garnet + biotite schist (samples EH45, EH46, and EH49) occur within and adjacent to leucosome material. \(X_{\text{Grs}}\) in most garnet porphyroblasts from these samples is nearly uniform at \(\sim 0.038\). One garnet from sample EH49 shows a very small and gradational shift to a slightly lower \(X_{\text{Grs}}\) mantle (Fig. 11f). Most grains show a slight \(X_{\text{Grs}}\) decrease along garnet rims to as low as \(\sim 0.031\).

Zoning patterns in garnet from paragneiss samples (EH30, EH31, EH37; Fig. 12a) resemble those found in garnet + biotite schist (EH43, EH46, EH49). Similar to the schist samples, \(X_{\text{Grs}}\) is nearly uniform (0.037–0.040) with a slight near-rim decrease to 0.035, consistent with diffusion during cooling. Paragneiss garnet is more common in melanosomes than in leucosomes or felsic bands, and contains inclusions of predominately quartz and biotite (some are poikiloblastic; Fig. 12c). Biotite inclusions in some grains preserve an earlier fabric that is discordant to the predominant fabric of the matrix (Fig. 12a).

Garnet from the migmatitic garnet + biotite schist (EH45, EH46, EH49) has relatively high-Y cores (250–400 ppm).

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**Fig. 9.** Partially resorbed garnet from migmatitic graphitic schist within the WLN (sample EH10). (a) Transmitted light thin-section scan of partially resorbed garnet shown in (b). The biotite + sillimanite + plagioclase corona texture surrounding the grain should be noted. (b) Ca X-ray map of garnet in (a) showing lower Ca garnet core, moderate Ca mantle, and higher Ca outer mantle zone. Low-Ca rims are embayed and only partially preserved. Higher Ca outer mantle is significantly higher than more moderate Ca garnet, and may represent garnet grown near peak pressure. (c) Digital scan of garnet shown in (d). Rectangle indicates the location of rutile inclusion discussed in text. (d) Ca map shows low-Ca core, higher Ca mantle that increases to a maximum, and a lower Ca rim that is truncated by resorption.
Fig. 10. X-ray maps of garnets from mylonitic paragneiss (EH21) from near the axis of the WLN. (a) Ca map of fractured, partially resorbed poikiloblastic garnet with preserved higher Ca ‘core’ and lower Ca zone. (b) Mn map shows preserved low-Mn core and high-Mn rim owing to significant garnet resorption. (c) Fe/(Fe + Mg) zoning follows that of Mn. (d) Ca map of staurolite-bearing (white arrow) garnet porphyroblast from Fig. 5c showing high-Ca core and low-Ca rim. White rectangle indicates rutile inclusion discussed in text. (e) Mn map of garnet shown in (d) showing rim increase in Mn. (f) Fe/(Fe + Mg) map of garnet in (d) and (e) shows a relatively thick Fe-rich rim, which may reflect a combination of growth and diffusion zoning.
Fig. 11. X-ray maps of garnet from migmatitic schist from the LBB: (a–c) sample EH46; (d–i) sample EH49. Values given are mole fractions except Y, which is by weight. (a) Fe/(Fe+Mg) map showing core–rim increase owing to retrograde diffusion. (b) Mn map showing uniform core and rimward increase. (c) P map showing zones of higher P along garnet rims. (Note high P along grain boundaries and fractures, as well as high-P plagioclase along the garnet rims.) (d) Fe/(Fe+Mg) map with lowest values in the garnet core. (e) Mn map showing variation in thickness of diffusion zoning, probably owing to variable amounts of garnet resorption along the rim. (f) Nearly uniform garnet Ca map with very minor lower Ca mantle. (g) Y map showing a significantly lower magnitude Y core compared with WLN garnet. Small, high-Y inclusions include xenotime, apatite, and monazite. (h) Sc map showing core–rim increase. (i) P map with partly preserved P annulus near the rim. Flushed out (white) inclusions in garnet are apatite. (Note high P along grain boundaries and fractures.)
and lower Y mantles (below electron microprobe detection limits). Some rims show a slight increase in Y as well, to as high as 145 ppm (Fig. 11g). High-Y cores contain both monzite and xenotime inclusions, although no xenotime is found in the low-Y mantles. Garnet cores may therefore have grown in equilibrium with xenotime. Minor xenotime also occurs along resorbed garnet rims. Scandium zoning is characterized by a relatively uniform low-Sc core (consistent with the higher Y core domain) with a broad increase through the mantle to the highest Sc values along some of the rims (Fig. 11h).

Phosphorus zoning is interesting in that near some rims where garnet is touching or nearly touching plagioclase, high-P zones occur (Fig. 11c). P in these zones is as high as 375 ± 33 ppm, whereas the rest of the garnet is nearly uniform with <100 ppm P. Very high P is observed along grain boundaries and fractures in samples EH46 and EH49 (Fig. 11c and i), and may indicate a late-stage P-bearing fluid was present. In sample EH49, P is not uniform in the high-P zone of garnet, but shows a gradual rimward increase with a peak before decreasing to low values at the rim (Fig. 11i). This P zoning may be related to partitioning with and/or breakdown of phosphate minerals (apatite?) during garnet growth, althoughapatite inclusions are present in the high-P zone (see Fig. 11i). Truncation of the high-P zone may indicate that all of a particular phosphate was consumed during this phase of garnet growth. Alternatively, the observation that high-P zones occur adjacent to plagioclase may indicate that P partitions into garnet during garnet growth and plagioclase breakdown, perhaps as a result of melting. In either case, the high-P zones could indicate garnet growth owing to melting reactions involving plagioclase and/or phosphates. A slight increase in the X_{Grs} content is coherent with the step in P, consistent with the anatectic growth of garnet.

Fig. 12. Textures and chemical zoning in LBB sample EH31. (a) Mg X-ray map of garnet and the surrounding matrix showing higher X_{Prp} core and low-X_{Prp} rim. A high-Mg biotite inclusion in garnet, shown by a circle, may be relict from an early phase of metamorphism. FM = Fe/(Fe + Mg). (b) Ca X-ray map showing plagioclase zoning in what may have been a garnet-bearing leucosome. (c) Crossed polarized light photomicrograph of poikiloblastic garnet porphyroblast with numerous quartz and biotite inclusions. (d) BSE image of area in white rectangle in (a) showing accessory phase inclusions in garnet that record prograde metamorphism.
Biotite
Biotite is abundant in the metapelitic samples of this study. The majority of matrix biotite from within the WLN that was analyzed shows generally only minor variation of Fe/(Fe + Mg) within each sample. Therefore it is inferred that retrograde (primarily net transfer) reactions probably have caused wholesale changes in matrix biotite Fe/Mg compositions.

Brown matrix biotite from the locally migmatitic graphitic schist unit (EH06, EH09, EH10) has Fe/[Fe + Mg] of 0.46–0.52 and 0.26–0.42 cations of Ti (on a 22 O basis).

In all WLN samples, biotite adjacent to resorbed garnet porphyroblasts is generally 0.02–0.05 lower in Fe/[Fe + Mg]. Green biotite shows a large variation of Ti content trending toward significantly lower values (<0.10).

This green biotite occurs as fine grains within garnet embayments or reaction coronas. The trend toward lower Ti contents for biotite along garnet rims is interpreted to represent retrograde growth of biotite at conditions below peak metamorphism. Variations in the Ti content of biotite along garnet rims seems loosely tied to the composition of the nearby garnet that is being resorbed. This may be due to the coincidence of rutile needles in garnet with higher X_{Grat}, suggesting a local kinetic effect on the availability of Ti to substitute into the biotite structure during retrograde garnet breakdown below rutile stability.

Biotite inclusions from the higher X_{Grat} mantle of a garnet porphyroblast in sample EH09 generally record similar to slightly lower Fe/[Fe + Mg] than matrix grains from this sample. This is probably a result of (1) biotite 'entrapment' occurring at higher grade conditions, followed by (2) retrograde exchange with the surrounding garnet resulting in an Fe/[Fe + Mg] shift toward that of the matrix biotite. Biotite inclusions are small (<30 µm wide) and complementary higher Fe/[Fe + Mg] garnet halos formed as a result of diffusion are very subtle and thin, perhaps in part owing to rapid cooling from higher temperature conditions.

In the LBB metapelites, biotite is brown, generally more Fe rich, and occurs in abundance within melanosome segregations. Matrix biotite from both schist and paragneiss is almost compositionally uniform within each analyzed sample, suggesting that biotite homogenized at some point on the P–T path, probably during retrograde metamorphism. Matrix biotite in locally migmatitic schists and paragneisses (EH30, EH33, EH45, EH46, EH49) has Fe/[Fe + Mg] of 0.61–0.69 and Ti of 0.25–0.42 cations per 22 O. A high-Ti biotite outlier in a felsic segregation of sample EH30, interpreted to represent leucosome, contains 0.47 Ti cations, and may be preserved biotite that grew during melt crystallization. A non-migmatitic paragneiss with almost no sillimanite (EH37) contains generally more magnesian biotite, with Fe/[Fe + Mg] of 0.52–0.54, which probably reflects very little retrograde metamorphism by retrograde net transfer reactions (ReNTRs).

Sample EH49 shows the widest range in biotite composition (see Electronic Appendix 4), and contains a number of biotite inclusions in garnet. Inclusion biotite must have undergone retrograde Fe/Mg diffusive exchange with garnet, yet preserves the highest Ti content for this sample (up to 0.47), probably representing relict prograde biotite grown at near peak temperatures. Biotite along garnet rims shows a variable composition that forms a trend from higher to lower values of both Ti and Fe/(Fe + Mg). This is probably due to biotite growth during continued breakdown of garnet along the retrograde path (e.g. Henry et al., 2005).

Feldspar
Within the WLN, plagioclase occurs throughout paragneisses in both mafic and felsic bands. In more mylonitic samples (e.g. EH21), felsic bands are almost exclusively composed of ribbon quartz. In the graphitic migmatitic schist unit, melanosome versus leucosome plagioclase shows different zoning patterns. Melanosome plagioclase has a lower X_{An} core (~0.33–0.36) with a slightly higher X_{An} rim overgrowth (~0.38–0.41; see Fig. 6c). Plagioclase formed in the leucosome (presumed to have initially crystallized from a melt) has higher X_{An} cores with slightly lower X_{An} rims. A 0.38 X_{An} plagioclase inclusion in garnet in sample EH21 occurs in the lower X_{Grat} rim domain observed in this sample, suggesting that this portion of the garnet equilibrated with the rims of matrix plagioclase, which are of a similar composition. Higher X_{An} (0.36–0.47) plagioclase forms reaction coronas around WLN garnet. X_{An} in the garnet corona varies as a function of the Ca content of the garnet that was replaced.

In the LBB, plagioclase from paragneiss samples shows more distinct preserved low-X_{An} cores (~0.17–0.23) and higher X_{An} rims (~0.23–0.29; e.g. Fig. 12b). Migmatitic garnet + biotite schist (samples EH46 and EH49) exhibits a range in plagioclase compositions near leucosome garnets. A core–rim relationship is not observed, and X_{An} in the matrix plagioclase ranges from 0.18 to 0.22. Plagioclase inclusions in garnet (e.g. Fig. 7b) are slightly higher in Ca (X_{An} = 0.24), suggesting that these grains were included in garnet prior to crystallization of the matrix grains. Plagioclase in sample EH45, which contains quartz veins but does not contain material interpreted as leucosome, is significantly lower in Ca, with X_{An} ranging from 0.13 to 0.16. This may indicate that some plagioclase produced on the prograde path did not react during partial melting, and melt extraction inhibited retrograde recrystallization or growth.

It is difficult to determine the stable mineral assemblage present when the plagioclase cores grew. Cores of zoned plagioclase from the WLN samples are interpreted to have grown on some portion of the prograde path. Plagioclase grains with lower X_{An} cores, assumed to have equilibrated with high-X_{Grat} garnet, and higher X_{An} rims may record a
significant decrease in pressure from a phase of higher $X_{Gr}$ garnet growth. The shift in plagioclase composition could reflect a period of plagioclase breakdown during anatexis on the decompression path followed by renewed growth with a higher $X_{An}$ content during melt crystallization. Leucosome plagioclase in migmatitic samples is interpreted to have crystallized from a melt phase during cooling, hence the core to rim anorthite content decreases slightly (observed in sample EH09).

Alkali feldspar occurs in paragneiss samples from the WLN as well as in migmatitic garnet + biotite schist from the LBB. Alkali feldspar from the WLN paragneiss is mainly interpreted to be metamorphic, produced by muscovite and possibly continuous biotite dehydration melting reactions. Metamorphic alkali feldspar analyzed is mainly $X_{Or}$ ~0.80–0.90, with rare slightly more sodic grains. Alkali feldspar is also present in many leucogranitic segregations throughout the crustal section, including leucosomes interpreted to have formed in situ in rocks from the LBB. In one leucogranite gneiss sample (EH35), $X_{Or}$ is limited to the higher values found in metamorphic grains, ~0.80–0.90.

**Rutile**
Rutile occurs in nearly every sample studied as inclusions in garnet. Most rutile inclusions are fine needles, although coarser anhedral rutile inclusions are also present (Fig. 5c). Rutile needles are generally absent from the low-$X_{Gr}$ cores and high-$X_{Gr}$ rims in WLN rocks, and probably represent exsolution of rutile from higher Ti garnet formed at relatively high pressures. WLN graphic schist samples also contain rutile inclusions in metastable kyanite porphyroblasts and/or metastable rutile in the matrix. Matrix rutile typically shows partial retrogression to ilmenite. Matrix rutile grains are considerably larger, suggesting that finer grains may have been fully replaced by ilmenite on the retrograde path. Some occurrences of matrix rutile are noted in LBB paragneiss EH30 along rims of matrix ilmenite. These rutile grains are rare and probably formed during retrograde exsolution from surrounding ilmenite.

Rutile was analyzed for zirconium by electron microprobe. X-ray mapping was performed to test for trace element ($Zr$, $Nb$, $Cr$, $Y$) zoning in the rutile that was analyzed for $Zr$. Trace element zoning in rutile inclusions in garnet is generally minimal. Rutile in contact with ilmenite grains shows an increase in Nb, perhaps owing to Nb diffusion driven by exclusion of Nb from the ilmenite structure during retrograde rutile breakdown.

**THERMOBAROMETRY**
**Major elements**
Independent reaction thermobarometric calculations performed on rocks from both the WLN and LBB are generally consistent with previous studies in the East Humboldt Range. However, coupled with zoning analysis, and guided by thermodynamic modeling (discussed below) of major melting reactions, prograde and retrograde compositions were used to estimate conditions at different points along a $P\cdot T$ path. Thermobarometric results were compiled graphically and are summarized in Fig. 13. Constraints on the maximum temperature at peak pressure and the peak temperature were estimated using garnet + biotite Fe/Mg exchange thermometry (Hodges & Spear, 1982) based on high-$X_{Gr}$ garnet mantles, Fe/(Fe + Mg) troughs, and garnet rims. Matrix biotite away from garnet was used in samples interpreted to have experienced ReNTRs. Pressures were estimated by garnet + aluminosilicate + plagioclase + quartz (GASP) barometry (Hodges & Crowley, 1985), garnet + hornblende + plagioclase + quartz barometry (GHPQ), and garnet + plagioclase + muscovite + biotite (GPMB) barometry (Hodges & Spear, 1982; Powell & Holland, 1988).

Retrograde exchange reactions (ReERs) and ReNTRs had a profound effect on the mineral compositions in this suite of samples. Retrograde exchange is evidenced by an increased $X_{phlogopitate}$ component in biotite and a sharp increase in Fe/(Fe + Mg) in garnet rims. ReNTRs have a similar effect on garnet zonning, although garnet is consumed by ReNTRs and shows enhyved edges, a reaction corona in the surrounding matrix, and truncated prograde zoning. ReNTRs have the opposite effect on biotite composition from ReERs, causing a shift in biotite composition from more phlogopite-rich to more annite-rich. Thermobarometry must account for retrograde changes in mineral chemistry and hence X-ray mapping is essential in interpreting the results of major phase thermobarometry (see Fig. 14). X-ray mapping of garnet (see Figs 8–12) was used to distinguish core domains from mantle troughs and diffusionally modified garnet rims. Thin-section scale variations in biotite chemistry were also recognized by X-ray mapping, and as a result biotite near garnet showing higher $X_{phlogopitate}$ was interpreted to have been affected by retrograde Fe/Mg exchange with garnet.

For a rock with little or no evidence of melting (EH37), peak temperature was estimated using garnet core compositions and matrix biotite away from garnets, interpreted to have not experienced ReERs. Because this sample contains neither leucosomes nor sillimanite (except for one retrograde textural occurrence), melting reactions that produce sillimanite probably did not occur.

Rocks in the upper limb of the WLN record a distinctly different history than rocks from the LBB. The retrograde assemblage suggests a similar exhumation path for these two blocks. Although the WLN rocks record higher pressure, the emplacement of the WLN may have occurred after partial decompression at high temperature and during significant anatexis. This requires not only burial
of Paleozoic sedimentary rocks in the WLN, but also high-temperature loading prior to decompression and emplacement in partially melted lower to middle crust (Fig. 13). Whatever event ‘decompressed’ these rocks from \( \sim 10 \) kbar must have begun prior to and probably continued during \textit{in situ} partial melting.

**Rutile thermobarometry**

Rutile in EHR metapelites most probably formed as a result of prograde breakdown of ilmenite in the presence of aluminosilicate and quartz to grow garnet via the GRAIL reaction

\[
\text{quartz} + \text{aluminosilicate} + \text{ilmenite} \rightarrow \text{almandine} + \text{rutile}
\]

Although this reaction is a well-calibrated geobarometer (Bohlen \textit{et al}., 1983), it requires knowledge of the reactive garnet and ilmenite compositions at the time of rutile growth, and hence application to rocks in which garnet zoning was modified by diffusion is problematic. Samples from the upper limb of the WLN preserve zoning that may be ‘reminiscent’ of prograde zoning, although garnet cores are largely homogeneous with respect to almandine, spessartine, and pyrope, suggesting modification by diffusion. Hence the \( X_{\text{Alm}} \) preserved in the cores of these garnets probably does not represent the initial garnet composition. Samples EH09 and EH10 contain abundant graphite, which acted to buffer the metamorphic fluid composition. Assuming that all fluid in the system is due to dehydration reactions, the maximum \( X_{\text{H2O}} \) is solely a function of temperature at fixed pressure for the graphite present system (Ohimoto & Kerrick, 1977; Connolly & Cesare, 1993). Hence, the \( /O_2 \) of the system is in effect buffered by the presence of graphite to be at close to common

![Fig. 13. Thermobarometry results. Reconstructed \( P-T \) paths are shown, as discussed in the text. Parallelisms represent geothermometer + geobarometer pairs calculated using measured compositional variation and calibrations discussed in the text. (a) WLN results and inferred \( P-T \) path. Peak pressure maxima indicate maximum temperatures for peak \( P \) based on high-\( X_{\text{Grs}} \) garnet and modified matrix biotite. Dashed \( P-T \) constraint (sample EH21) is based on Zr-in-rutile thermometry (shown as gray bands for pre- and post-high-\( X_{\text{Grs}} \) garnet growth). Dashed solidus is from thermodynamic modeling based on sample EH10 and applies to rocks of a similar pelitic bulk composition. (b) LBB thermobarometry results as in (a). Dashed solidus is from thermodynamic modeling based on sample EH49.](http://petrology.oxfordjournals.org/)

![Fig. 14. Mg X-ray map of garnet and surrounding matrix in LBB sample EH49. White circles are electron probe spots on garnet and biotite with FM = Fe/(Fe + Mg) given. Fe/Mg exchange thermometry results are given [calibration of Hodges & Spear (1982)], demonstrating the effects of (1) retrograde Fe/Mg exchange to record retrograde, sub-peak temperatures along garnet rims and (2) retrograde and net transfer reactions to record false temperatures above the thermal peak.](http://petrology.oxfordjournals.org/)

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crustal values at lower temperatures and increasingly reducing relative to common $O_2$ buffers (e.g. magnetite-
wustite, $R_eO_3 = R_eO + \frac{1}{2}O_2$) with higher temperatures. Therefore, ferric iron is minimal in ilmenite and garnet
(as suggested by calculation from stoichiometric assumptions; see Electronic Appendix 6) in these rocks. GRAIL
barometry using the modified garnet cores and pure ilmenite yields unreliable pressures of $\sim 5$–6 kbar.

The interpretation that significant garnet growth in the WLN rocks took place in the kyanite field, partially owing
to the breakdown of staurolite, and the observation that rutile occurs as inclusions in garnet, suggests that some portion of the garnet porphyroblasts equilibrated with rutile and ilmenite in the kyanite field. No evidence of sillimanite
before kyanite is found in these rocks. Therefore, rutile probably formed as part of the prograde assemblage garnet +
kyanite + quartz + plagioclase + ilmenite + rutile + zircon ($\pm$ muscovite $\pm$ biotite). Using calculated GRAIL equilibria,
limited to the kyanite field, with pure ilmenite in equilibrium with the modified garnet composition requires
rutile growth at temperatures lower than $\sim 600^\circ$C. Clearly, garnet continued to grow following rutile crystallization
and GRAIL equilibration, as the majority of the rutile preserved in these rocks is as inclusions in garnet. Therefore,
the garnet in the GRAIL assemblage may have been effectively cut off from the matrix during continued garnet
growth, and subsequently altered by major cation volume diffusion (in sample EH10 to lower $X_{Alm}$ values) at elevated
temperatures.

Zirconium-in-rutile

The Zr-in-rutile thermometer (Zack et al., 2004; Watson et al., 2006; Tomkins et al., 2007) was applied to rutile
inclusions in garnets and matrix rutile from both the WLN and the LBB (see Table 1; Fig. 15). X-ray maps of Zr, Nb, and
Fe indicate minor zoning of these trace elements in some rutile grains (Fig. 15b, c, and e). Partial rutile replacement
by ilmenite is recognized in matrix grains or grains connected to the matrix via cracks in enclosing garnet
porphyroblasts. Ilmenite ‘stringers’ and decorations on rutile grains are common and probably represent both exsolution
of ilmenite from rutile as well as replacement of rutile by ilmenite during retrograde decompression (Fig. 15d and f).

Zr-in-rutile temperatures (Table 1) were calculated using the calibration of Tomkins et al. (2007) and range from
$< 500^\circ$C in matrix grains to $722^\circ$C for a rutile inclusion in garnet. The majority of rutile occurs as inclusions in
garnet or kyanite. Rutile inclusions in garnet or kyanite give temperatures from 675 to $722^\circ$C for rocks within
the WLN (Fig. 15a, e, and f). In the LBB, rutile inclusions in garnet give temperatures from 655 to $700^\circ$C (Fig. 15g
and h).

In the WLN samples, rutile occurs mainly as inclusions in garnet mantles, suggesting that significant garnet
growth proceeded after rutile crystallization and equilibration with zircon. It is possible that the Zr-in-rutile
temperatures recorded by rutile inclusions in the garnet mantle represent the actual temperature of rutile growth.
If the temperatures do reflect growth, then the GRAIL barometer must have been grossly altered by dif-
fusional alteration of the garnet core composition to significantly lower $X_{Alm}$ values, giving an erroneous lower estimated pressure, well within the sillimanite field. This would also imply that the rutile + zircon + quartz assemblage remained open to higher temperatures, resulting in significant diffusion of Zr into rutile during slow heating. Alternatively, if rutile growth took place at the temperatures recorded by the Zr thermometer, then slow heating is not necessary to explain the homogeneous Zr content of the rutile inclusions. Thermodynamic modeling of sample EH10 suggests that rutile stability is reached in this case at $\sim 675^\circ$C and $\sim 7.5$ kbar, and hence GRAIL barometry does not represent equilibrium.

The Zr-in-rutile temperatures provide an important constraint on the temperatures of garnet growth periods
in the WLN. Rutile inclusions in garnet in samples EH09 and EH10 occur within the higher $X_{Grs}$ mantle zone
(Fig. 9f). Although no rutile is found in the low-$X_{Grs}$ core, the interpretation that the garnet core grew below rutile
stability is preferred. The garnet mantle grew during and/or after equilibration of the rutile inclusions with
matrix zircon and quartz. Therefore the Zr-in-rutile temperatures, assuming the system closed to rutile + matrix
equilibration during garnet growth, represent a minimum for garnet mantle growth. Although the temperatures
are below estimates of peak temperature based on melting reaction textures, they suggest that most garnet mantle
growth occurred along a steep P–T path segment (nearly isothermal loading) above the Zr-in-rutile temperature.
Assuming that the lower $X_{Grs}$ core grew in equilibrium with plagioclase and quartz, the garnet mantle records an
increase in the $X_{Grs}$ component of 0.025–0.03, which corresponds to as much as 2–3 kbar of loading assuming equilib-
rium with plagioclase cores.

In WLN paragneiss (EH21), a rutile inclusion occurs in a distinctly different garnet domain from those in EH09
and EH10 (see Figs 10d and 15f). The lower $X_{Grs}$ garnet surrounding this higher Zr rutile is interpreted to be garnet
grown during biotite dehydration melting. The rutile inclusion is significantly higher in Zr ($\sim 766$ ppm) and gives a
temperature of $722 \pm 6$–7°C. Another inclusion in the same garnet domain gives a consistent temperature of
$715 \pm 4$°C. There is no evidence to suggest that these compositions have been subsequently modified by diffusive ex-
change with garnet, but they may have experienced diffusional homogenization since being enclosed in garnet.
In the case of diffusional homogenization, the Zr-in-rutile results from EH21 represent a minimum temperature for
the prograde growth of the surrounding garnet.
### Table 1: Zr in rutile analyses from East Humboldt range metapelites

<table>
<thead>
<tr>
<th>Sample Grain Context</th>
<th>n*</th>
<th>Av. Zr (ppm)</th>
<th>1σ SD† (ppm)</th>
<th>exp. SD‡ (ppm)</th>
<th>1σ SE (ppm) at 8 kbar</th>
<th>2σ SE (°C) at 8 kbar</th>
<th>2σ SE (°C)</th>
<th>Notes</th>
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<tbody>
<tr>
<td>Winchell Lake nappe</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>EH09b-p matrix</td>
<td>4</td>
<td>397</td>
<td>56</td>
<td>18</td>
<td>—</td>
<td>2/25</td>
<td></td>
<td></td>
</tr>
<tr>
<td>EH09a-p incl in grt</td>
<td>4</td>
<td>452</td>
<td>29</td>
<td>18</td>
<td>—</td>
<td>11/11</td>
<td></td>
<td></td>
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<tr>
<td>EH10b-p incl in ky</td>
<td>1</td>
<td>490</td>
<td>21</td>
<td>18</td>
<td>—</td>
<td>663</td>
<td></td>
<td></td>
</tr>
<tr>
<td>EH21a-p incl in grt</td>
<td>1</td>
<td>766</td>
<td>27</td>
<td>19</td>
<td>—</td>
<td>14</td>
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<tr>
<td>Lizzies Basin block</td>
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<td></td>
</tr>
<tr>
<td>EH30b-p incl in grt</td>
<td>4</td>
<td>566</td>
<td>31</td>
<td>18</td>
<td>—</td>
<td>688</td>
<td></td>
<td></td>
</tr>
<tr>
<td>EH37b-p incl in grt</td>
<td>4</td>
<td>601</td>
<td>38</td>
<td>18</td>
<td>—</td>
<td>606</td>
<td></td>
<td></td>
</tr>
<tr>
<td>EH45a-p incl in grt</td>
<td>4</td>
<td>491</td>
<td>24</td>
<td>18</td>
<td>—</td>
<td>693</td>
<td></td>
<td></td>
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<tr>
<td>EH46b-p incl in grt</td>
<td>4</td>
<td>484</td>
<td>23</td>
<td>18</td>
<td>—</td>
<td>681</td>
<td></td>
<td></td>
</tr>
<tr>
<td>EH49a-p incl in grt</td>
<td>4</td>
<td>468</td>
<td>34</td>
<td>18</td>
<td>—</td>
<td>678</td>
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<tr>
<td>EH49b-p incl in grt</td>
<td>4</td>
<td>493</td>
<td>24</td>
<td>18</td>
<td>—</td>
<td>682</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* n*, number of spots; total analyses = n × 4.  
† Measured standard deviation for each grain.  
‡ Measured standard deviation propagated to temperature.  
§ Measured standard deviation based on counting statistics for analytical protocol.  
¶ Measured standard error of all temperatures for the grain, given where expected SD ≥ measured SD.

Analytical protocol is summarized in Electronic Appendix 1.
Thermodynamic modeling considerations suggest that the temperatures of anatectic garnet growth may have been slightly higher than this value, consistent with some diffusional modification. Based on this interpretation, the high-$X_{Grs}$ garnet core (mantle in EH09 and EH10) must have grown at a temperature below ~740°C. Therefore, Zr-in-rutile thermometry for these samples (EH09, EH10, and EH21) indicates that a phase of tectonic loading and

Fig. 15. Rutile from the East Humboldt Range: (a–f) are from WLN; (g, h) are from LBB. White bands labeled ‘trav’ represent electron microprobe traverses. The average Zr content and corresponding temperature are given for each grain. (a) Backscattered electron image (BSE) of matrix rutile from sample EH09 showing Zr analysis locations and results. (b) Nb X-ray map of rutile in (a). (c) Zr X-ray map of rutile in (a). (d) BSE image of rutile and zircon inclusions in garnet from Fig. 9. (e) Zr X-ray map of rutile in white box in (d). (f) BSE image of rutile inclusion from Fig. 10d. (g) Rutile inclusion in garnet with biotite from sample EH46. (g) Rutile inclusion in garnet from sample EH37.
subsequent decompression occurred in the WLN between \(\sim 675\) and \(\sim 740^\circ C\).

**THERMODYNAMIC MODELING**

Relict phases, garnet zoning and reaction texture observations, combined with thermobarometry, are better interpreted in conjunction with appropriate equilibrium assemblage diagrams contoured for phase composition and abundance. Calculations were performed using the thermodynamic data of Holland & Powell (1998), White et al. (2001), the database of F. S. Spear et al. (unpublished data), and a combination of program Gibbs (Spear & Menard, 1989; Spear et al., 2001) and Theriaq-Domino (de Capitani & Brown, 1997; de Capitani & Petrakakis, 2010).

Appropriate generalized bulk compositions in the system MnNCKFASH (± Ti) were calculated with program Gibbs from estimated volumetric proportions of peak phases and measured compositions of solution phases. As EHR migmatites show evidence of melt migration and segregation from the reactive bulk composition, additional melt in equilibrium with modeled solidus phases was added, using the melt reintegation method of Indares et al. (2008). A melt composition was calculated at the muscovite dehydration melting reaction for the proposed \(P-T\) path, at 748°C, 8.6 kbar for a model of WLN sample EH10, and at 722.5°C, 7.5 kbar for a model of LBB sample EH49. This melt was then added to the system until the effective muscovite dehydration melting reaction shifted to \(H_2O\)-saturated conditions (equivalent to 10 mol % melt for EH10, 13 mol % melt for EH49). This does not significantly change the temperature of the effective solidus (interpreted as the muscovite-out–alkali feldspar-in reaction) at the expected pressure range for melting, but it does affect the stability fields of a number of subsolidus phases interpreted to have been part of the prograde reaction history.

Composite equilibrium assemblage models were produced (Figs 16 and 17) with the melt-reintegrated composition, using a fixed \(H_2O\) content based on the reintegrated melt composition for the suprasolidus portion (including the solidus), and an \(H_2O\)-saturated composition for all of the subsolidus fields. In subsolidus fields, \(H_2O\) is treated as a fractionated or isolated phase and hence this portion of the model was calculated using a melt-free database. Calculations were performed, and equilibrium assemblage diagrams were constructed, with Theriaq-Domino. A number of assumptions are required to apply thermodynamic modeling to the interpretation of petrological observations. This method does not account for (1) fractional crystallization, (2) subsolidus fluid flux into or out of the system, (3) the complete chemical composition including \(Fe^{2+}\) and minor elements, (4) possible multiple or protracted melt loss events and (5) incorporation of Fe and Mg into the melt. Equilibrium assemblage diagrams are therefore presented as an illustration of the interpretation of petrological observations for the WLN. In particular, calculated isopleth maps are not taken as accurate with respect to absolute compositional predictions, but rather represent guides with approximations of equilibrium chemical gradients in \(P-T\) space.

Both melt-reintegrated models yield peak pressure and/ or peak temperature assemblages that are consistent with petrographic observations and major phase thermobarometry with one exception. Alkali feldspar is notably absent from the WLN migmatitic schist (EH05–EH10). It is possible and likely that all alkali feldspar broke down owing to a retrograde net transfer reaction during melt crystallization, the same reaction that resulted in significant garnet resorption and production of biotite and sillimanite. LBB migmatitic schist samples (EH45, EH46, EH49) do contain alkali feldspar, interpreted to have been part of the peak metamorphic assemblage.

**Modeled solution phase compositions**

The prograde, pre-melt extraction model presented in Fig. 16 predicts two major phases of garnet growth in the subsolidus region and minor additional supersolidus garnet growth for the WLN migmatitic schist (EH10). According to the model, garnet initially began to grow owing primarily to the breakdown of chlorite, followed by little growth or slight resorption at the staurolite-in reaction (Fig. 16b). Garnet growth resumed in a staurolite-bearing assemblage and grew as a result of staurolite breakdown and a continuous reaction in a subsolidus kyanite-bearing assemblage to peak pressure and \(X_{Sps}\) content (Fig. 16b and c). At this point decomposition accompanied resorption, although very little change in the \(X_{Sps}\) content is depicted (Fig. 16d). Crossing the muscovite dehydration melting reaction results in some anatectic garnet growth to peak temperatures, followed by resorption during cooling and melt crystallization. Along the \(P-T\) path, major resorption is predicted to occur on steeply decompressional segments below the solidus, and shallower cooling segments in melt-bearing assemblages.

The EH10 (WLN) garnet composition during growth along the proposed \(P-T\) path is initially \(X_{Sps}\)-rich and with moderate \(X_{Grt}\) (see Fig. 16c and d). Garnet interpreted to have grown along the prograde \(P-T\) segment below staurolite stability (garnet zone) may not have been in equilibrium with plagioclase, as the plagioclase present may have been detrital and not entirely reactive, acting to sequester Ca from the growing garnet. More moderate \(X_{Sps}\) garnet is predicted to grow in the presence of staurolite, followed by relatively low-\(X_{Sps}\) garnet in the kyanite-bearing field. Garnet grown in the kyanite field along the steep loading segment of the \(P-T\) path will become increasingly \(X_{Grt}\)-rich. Finally, anatectic garnet is significantly lower in \(X_{Grt}\) than subsolidus garnet grown near peak pressures. Model equilibrium \(X_{An}\) compositions...
Fig. 16. Equilibrium assemblage diagrams in the MnNCKFMASH system constructed using thermodynamic data of Holland & Powell (1998) and the melt model of White et al. (2001). (See Supplementary Data Electronic Appendix 3 for activity models and Table 2 for bulk compositions used.) L, liquid (melt). (a) Assemblage stability fields for prograde composition based on WLN sample EH10 with reintegrated melt. Subsolidus region is entirely vapor saturated, using an H2O-enriched bulk composition. (See text for discussion.) Dot pattern indicates peak assemblage. Shading darkness correlates to assemblage variance. Dashed thick gray line with arrow represents inferred \( P-T \) path, labeled with reaction numbers from text. Field labels: (A) Bt + Grt + Ms + Qtz; (B) Chl + Grt + Ms + Pl + Qtz; (C) Chl + Grt + Ms + Pl + Qtz + St; (D) Bt + Grt + Ms + Pl + Qtz + St; (E) Bt + Grt + pl + Qtz + Sil + Ld; (F) Cdr + Grt + Pl + Qtz + Sil + Ld; (G) Cdr + Grt + Pl + Sil + Ld; (H) Chl + Grt + Ms + Pl + Qtz; (I) Bt + Cdr + Grt + Pl + Qtz + Sil + Ld; (J) Bt + Chl + Cld + Pl + Qtz; (K) And + Bt + Cdr + Kfs + Pl + Qtz.

(b) As in (a) with volume per cent isopleths of garnet. Some isopleths are omitted for clarity. Black \( P-T \) path segments represent garnet growth; white segments are garnet resorption. (c) As in (a) with isopleths of \( X_{\text{Grt}} \). (d) As in (a) with isopleths of \( X_{\text{Pl}} \). (e) As in (a) with isopleths of \( X_{\text{An}} \), in plagioclase. Some isopleths are omitted for clarity.
Fig. 16. Continued.
are consistent with plagioclase growth during a higher pressure metamorphic phase in the subsolidus region (Fig. 16c), overgrown by plagioclase during retrograde melt crystallization following melt extraction or loss.

Garnet in the EH49 (LBB) model is inferred to have been almost completely resorbed in the presence of staurolite, suggesting that garnet may have re-nucleated during staurolite breakdown (Fig. 17a). Modeled isopleths of Fe/(Fe + Mg) in garnet (Fig. 17b) are relatively low at estimated peak conditions, suggesting that wholesale changes in the measured Fe/(Fe + Mg) are dramatically affected by diffusion on the retrograde path. The proposed prograde $P^T$ path through modeled plagioclase compositions spans a wide range in $X_{An}$ (Fig. 17c), which is difficult to interpret without knowing whether plagioclase was initially detrital or grew during a phase of prograde metamorphism. It is likely that measured matrix plagioclase compositions ($X_{An} = 0.18–0.22$ for EH49) entirely reflect growth during retrograde melt crystallization, following extraction of a significant volume of melt.

**MAJOR PHASE REACTION HISTORY**

Reactions and textures deduced from petrological analysis are consistent with the results of thermodynamic modeling. An interpreted reaction history is presented, with reference to the equilibrium assemblage diagram for a WLN rock (Fig. 16a) and an LBB rock (Fig. 17a). Reaction numbers presented in the text are linked to Figs 16a and 17a.

**Winchell Lake nappe**

In the WLN graphitic schist (EH06, EH09, EH10), some garnets contain a relict core, in some cases rich in quartz inclusions, which grew during prograde garnet zone metamorphism, most probably by the breakdown of chlorite and chloritoid owing to the garnet-in isograd reaction

$$\text{chlorite} + \text{chloritoid} + \text{quartz} = \text{garnet} + \text{H}_2\text{O}.$$  \hspace{1cm} (2)

Garnet cores do not appear to be resorbed, as euhedrally shaped trace element (Y, Sc) zoning is observed, probably representing the disappearance of xenotime from the rock (e.g. Pyle & Spear, 1999). With further heating, some garnet growth would have continued via the staurolite-producing reaction

$$\text{chloritoid} + \text{muscovite} + \text{quartz} = \text{garnet} + \text{staurolite} + \text{biotite} + \text{H}_2\text{O}.$$  \hspace{1cm} (3)

In lower aluminum WLN pelites, garnet growth probably began by the reaction

$$\text{chlorite} + \text{muscovite} + \text{quartz} = \text{garnet} + \text{biotite} + \text{H}_2\text{O}.$$  \hspace{1cm} (4)
**Fig. 17.** Equilibrium assemblage diagrams in the MnNCKFMASH system. L, liquid (melt). Dashed thick gray or black line with arrow represents inferred $P^T$ path. (See Electronic Appendix 5 for activity models and Table 2 for bulk compositions used.) (a) Assemblage stability fields for pro-grade composition based on LBB sample EH49 with reintegrated melt. Subsolidus region is entirely vapor saturated, using an H$_2$O-enriched bulk composition. (See text for discussion.) Dot pattern indicates peak assemblage. Shading darkness correlates to assemblage variance. $P^T$ path is labelled with reaction numbers from text. Field labels: (A) Bt + Grt + Ms + Qtz; (B) Bt + Grt + Kfs + Ky + Pl + Qtz + L; (C) Grt + Kfs + Qtz + Sil + L; (D) Bt + Ms + Pl + Qtz + St; (E) Bt + Chl + Ms + Pl + Qtz + St; (F) And + Bt + Grt + Ms + Pl + Qtz; (G) Bt + Card + Kfs + Pl + Qtz + Sil + L; (H) And + Bt + Card + Kfs + Pl + Qtz; (I) And + Card + Kfs + Pl + Qtz + L; (J) And + Card + Kfs + Pl + L. (b) As in (a) with Fe/(Fe+Mg) isopleths for garnet. (c) As in (a) with isopleths of $X_{An}$ in plagioclase. Some isopleths are omitted for clarity.  

(continued)
Garnet growth would be continuous via reaction (4). No chloritoid inclusions are found in garnet in rocks for which abundant aluminosilicates reflect an aluminous bulk composition (i.e. Fig. 16).

Staurolite growth most probably led to the removal of chlorite by the reaction

\[
garnet + \text{chlorite} + \text{muscovite} = \text{staurolite} + \text{biotite} + \text{quartz} + \text{H}_2\text{O}
\] (5)

which was probably the main staurolite-producing reaction. Significant garnet breakdown by this reaction may have occurred in some rocks, although it is not reflected in chemical zoning profiles. Garnet consumption probably took place above xenotime stability, based on a euhedral shape to the high-Y zone (Fig. 8d) that shows no evidence of resorption.

Garnet growth resumed during the breakdown of staurolite by a reaction that may have produced large kyanite porphyroblasts that are present in samples EH09 and EH10:

\[
\text{staurolite} + \text{muscovite} = \text{garnet} + \text{biotite} + \text{kyanite} + \text{quartz} + \text{H}_2\text{O}
\] (6)

Garnet growth probably continued to peak pressure conditions above the staurolite field by the continuous reaction

\[
\text{biotite} + \text{kyanite} + \text{quartz} = \text{garnet} + \text{muscovite}
\] (7)

It is possible that most kyanite was completely consumed by reaction (7) to grow garnet. Alternatively, kyanite may have been replaced by sillimanite during a decompression segment of the prograde path that passed into the sillimanite field.

At the highest pressure recorded in these rocks, reaction (7) was active along with the GASP reaction

\[
3\text{ anorthite} = \text{kyanite} + \text{quartz} + \text{grossular}
\] (8)

which together resulted in significant \(X_{\text{Grs}}\)-rich garnet growth.

Decompression, starting in the kyanite field, resulted in the partial resorption of garnet via the retrograde forms of reactions (7) and (8). Theoretically, kyanite and biotite growth would have occurred during this initial decompression phase, prior to reaching dehydration melting reactions and/or the sillimanite field. Although nearly all matrix kyanite in samples EH05–EH10 is partly replaced by sillimanite, it is uncertain whether the relict kyanite had grown during initial decompression. The samples with relict kyanite also contain abundant large knots of fibrous sillimanite and prismatic sillimanite grains. Some severely resorbed relict kyanite is completely rimmed by plagioclase with no evidence of replacement by sillimanite (see Fig. 6a). This rim may have formed as a result of increased \(\text{Ca}^{2+}\) and \(\text{Na}^+\) intergranular diffusion at near peak metamorphic conditions, perhaps related to growth of higher

\[\text{Fig. 17. Continued.}\]
The presence of melt occurs as embayed garnet rims showing diffusional Fe/(Fe + Mg) zoning, as well as artificially high garnet core + biotite Fe/Mg temperatures owing to melt-involving retrograde net transfer reactions that produced $X_{\text{phlogopite}}$-rich biotite. Alkali feldspar is not present in these rocks, suggesting that it was entirely consumed by the retrograde form of reaction (10):

$$\text{garnet} + \text{alkali feldspar} + \text{melt} = \text{biotite} + \text{plagioclase} + \text{sillimanite} + \text{quartz.} \quad (10r)$$

This retrograde net transfer reaction significantly altered the bulk mineralogy of these metapelitic rocks. The absence of alkali feldspar and muscovite in these rocks confirms that (1) significant melt was separated from the reactive portion of these rocks above the solidus and (2) all of the remaining melt and alkali feldspar after melt separation reacted with garnet to grow the products of reaction (10r) that dominate the melanosomes of these migmatites. Continuous biotite dehydration melting, reaction (10), acting to grow anatectic garnet may have been minimal in WLN rocks, although melt crystallization owing to reaction (10r) consumed significant quantities of garnet, including nearly all anatectic garnet, to produce retrograde biotite (Fig. 6f). The progress of this reaction was probably limited by the amount of melt and alkali feldspar available to react, and hence $P-T$ conditions under which reaction (10r) ceased are related to the bulk composition of the rock.

### Lizzies Basin Block

In the LBB, no evidence is found for the early phase of higher pressure, kyanite-grade metamorphism. One interpretation of this observation is that the LBB did not experience the higher pressure conditions and hence the emplacement of the WLN is an important phase that acted to juxtapose rocks of distinctly different parageneses. Alternatively, if the LBB experienced the same higher
pressure metamorphism, complete overprinting and/or diffusional resetting of the original mineral zoning would have been required to erase all record of this event. If this were the case one might expect similar zoning patterns for trace elements and $X_{\text{Grs}}$ in garnet from both the WLN and LBB, although garnet preserves different trace element and $X_{\text{Grs}}$ zoning for the two blocks.

Initial garnet growth in migmatitic pelites from the LBB probably occurred during prograde reactions involving the breakdown of hydrous phases [chlorite + biotite; e.g. reaction (3)]. Although no evidence is preserved for garnet resorption, thermodynamic modeling results suggest that it is possible that some garnet was resorbed during staurolite growth. Partial melting in the sillimanite field broke down muscovite by reaction (9) to produce significant melt (Fig. 7a) as well as abundant fibrolitic sillimanite and alkali feldspar present in melanosomes. Biotite dehydration melting with garnet and alkali feldspar growth then continued and additional melt was produced by continuous reaction (10). Some garnet shows a rimward decrease in the inclusion density (Fig. 7b), interpreted to represent the start of a new phase of garnet growth, possibly during continuous melting reaction (10).

Based on (1) the relatively large estimated modal amount of melanosome or mesosome material on hand-sample scale (~80–90%), (2) the relatively low modal amount of muscovite and its cross-cutting nature with respect to biotite and sillimanite (Fig. 7d), and (3) the observation of interconnected web-like melt textures (Fig. 4f), melt was most probably removed from the reactive bulk composition by some combination of isolation and extraction during deformation near peak conditions. As discussed above, and interpreted by previous workers (Wright & Snoke, 1993; Batum, 1999; McGrew et al., 2000; McGrew & Snoke, 2010), some proportion of ~1–10 m scale leucogranitic dikes may represent coalesced melt produced during prograde metamorphism and effectively separated from the metapelitic layers, driving them toward a ‘restitic’ composition.

Although a significant portion of the melt phase was lost as a result of mechanical removal, much of the remaining melt back-reacted with the surrounding rock to break down prograde garnet and alkali feldspar (Fig. 7b, c and e) via reaction (10r). This is observed as embayed garnet grains and biotite + sillimanite pseudomorphs after garnet (Fig. 7f). Further retrograde garnet breakdown below the effective solidus probably continued via the anhydrous reaction
garnet + muscovite = biotite + sillimanite + quartz (11)

which, in addition to melt crystallization reactions, helped produce abundant biotite and sillimanite in these rocks. Isopleths of garnet abundance as a result of this reaction are relatively shallow in slope ($\Delta P/\Delta T$; e.g. Spear et al., 1999).

Therefore, the function of this retrograde reaction represents a continued component of decompression below the effective solidus. Reaction (11), in conjunction with Fe/Mg exchange with biotite on the subsolidus retrograde path, results in steep diffusional increases in Fe/(Fe + Mg) near LBB garnet rims, suggesting that low-temperature retrograde cooling may have been relatively slow.

**DECOMPRESSION OF THE WINCHELL LAKE NAPPE**

Structural evidence for the early phase of decompression from the inferred WLN $P$–$T$ path (Fig. 13) is not recognized in the northern East Humboldt Range. Several possible explanations exist for the tectonic cause of this decompression event: (1) erosion; (2) vertical ductile thinning; (3) normal faulting; or potentially some combination of these three (see Ring et al., 1999). Erosion as a mechanism for decompression is possible, perhaps as a response to thrusting, although Late Cretaceous to Eocene strata are relatively sparse across the Sevier hinterland (Vandervoort & Schmitt, 1990), suggesting that erosion played only a minor or supporting role in decompression.

Vertical ductile thinning in the lower to middle crust is another mechanism of decompression that is compatible with significant crustal thickening by thrusting and nappe formation in the Sevier hinterland (e.g. Ring & Kassem, 2007). Subhorizontal foliations in the WLN (and LBB) are consistent with vertical ductile thinning as a result of deformation during high-grade metamorphism (Kassem & Ring, 2004). However, decompression by vertical ductile thinning alone is not compatible solely with the phase of emplacement of the WLN onto the LBB for two reasons. First, the LBB records somewhat higher peak temperatures (~30° higher than the WLN), which, based on inferred retrograde $P$–$T$ paths, occurred prior to juxtaposition of the two blocks, suggesting that some cooling and exhumation of the LBB predated WLN emplacement. Second, the record of nearly isothermal loading followed by ~3–5 kbar decompression with minor heating of the WLN is not observed in the LBB, suggesting that these two blocks were well separated during WLN loading and early WLN decompression.

Normal faulting in the middle to upper crust is a possible explanation for significant decomposition. The nearby Pequop fault has been interpreted to have accommodated ~10 km of crustal thinning in the Late Cretaceous (Camilleri & Chamberlain, 1997). This structure was rooted beneath the RM–EHR and therefore could not have accommodated the WLN decomposition. Although no preserved Late Cretaceous normal fault system is observed in the present configuration of the RM–EHR, it is perhaps possible that a localized system similar to the Pequop fault could have been active at shallower crustal
levels above a decompressing WLN. Syncontractional normal faulting is a very important exhumation mechanism for the High Himalayan Crystalline complex (Burchfiel & Royden, 1985; Burchfiel et al., 1992; Grujic et al., 1996; Ring & Glosny, 2010). Although no evidence exists for a major structure such as the South Tibet Detachment system in the Sevier orogen, gravitational collapse of the Sevier hinterland may have been accommodated by localized extensional deformation synchronous with foreland shortening. Normal faulting in the shallow crust could have a significant effect on the exhumation history of the middle to lower crustal section. Alternatively, the formation of an extensional allochthon in the lower to middle crust (e.g. Hodges & Walker, 1992) could accommodate some decompression, although this creates a significant space problem with no evidence for extrusion to the surface.

The driving force for, and the significance of, normal faulting to accommodate syncontractional exhumation in a portion of the crustal section is a topic of much debate in the literature (e.g. Hodges & Walker, 1992; McGrew et al., 2000; Wells & Hoisch, 2008; Druschke et al., 2009b; Long, 2012; Miller et al., 2012; Wells et al., 2012). A model proposed by Wells & Hoisch (2008) and Wells et al. (2012) argues that mantle delamination occurred in the Late Cretaceous (~75–67 Ma) beneath the Sevier hinterland, resulting in cryptic Late Cretaceous syncontractional exhumation and leucogranite magmatism, in part as a response to localized changes in gravitational forces in an orogenic wedge. Although this model explains anatexis and a potential change in the thermal structure of the lower crust to explain heating during WLN decompression, it also predicts (1) significant extension in the upper crust, and (2) a juvenile component to peraluminous granitoid magmatism (Wells et al., 2012). As mentioned above, Late Cretaceous extensional structures are poorly preserved. Late Cretaceous magmatism in the RM–EHR includes no juvenile component, and ages for peraluminous leucogranites span a much larger range (Howard et al., 2011) than the proposed delamination event, suggesting different or additional triggers for crustal anatexis.

**CONCLUSIONS: THE SIGNIFICANCE OF PARTIAL MELTING**

In the WLN, peak temperature conditions (~760°C) are estimated to be higher than the temperature at peak pressure (~740°C), suggesting that heating continued during early decompression. The source of heat during decompression is difficult to pinpoint and may be related to advection of heat from the lower crust by leucogranite magmatism in the Late Cretaceous (Wright & Snoke, 1993; McGrew et al., 2000). Although estimated peak conditions in the WLN involved widespread partial melting, it is not straightforward to link the abundant leucogranite dikes and sills interpreted as Late Cretaceous to evidence of melting in metapelitic rocks. Although McGrew et al. (2000) argued that leucogranite intrusions are associated with WLN pelites, it cannot be demonstrated that all leucogranite magmatism was contemporaneous with local in situ melt crystallization. There is a possibility that some (or most) leucogranite intrusion and melt migration from deeper crustal levels during deformation may pre-date in situ melting related to decompression, and acted as a ‘heat source’ to push the WLN P–T path into the melt field.

In metapelitic rocks from the LBB, partial melting occurred during prograde metamorphism, directly in the sillimanite field. The absence of relict kyanite and staurolite plus differences in the XGrs and trace element zoning in garnet porphyroblasts in migmatic rocks from each block supports this interpretation. The different P–T history for LBB rocks indicates that the emplacement of the WLN is an important event in the tectonic history of the RM–EHR core complex that acted to juxtapose two blocks with different petrogenetic and melting histories. The two blocks share a similar structural style and similar constraints on retrograde metamorphic re-equilibration. The LBB does not show an episode of nearly isothermal loading, which would have occurred had the WLN been emplaced prior to peak temperatures (see Spear et al., 2002). Therefore, juxtaposition of these metamorphosed blocks must have occurred after, but probably near, the thermal peak of metamorphism, consistent with the interpretations of the timing of the WLN formation of McGrew et al. (2000). The results of this study imply a single metamorphic episode in the WLN (and Clover Hill) resulting in continued heating through peak pressure and then peak temperature phases of metamorphism.

**In situ anatexis in the WLN and the subsequent emplacement of the WLN must have occurred after a phase of decompression with heating that is recorded only by WLN rocks. In contrast, anatexis of LBB metapelites occurred during a pressure increase along a P–T segment with a positive slope across dehydration melting reactions. Despite melting at similar pressures, the difference in these two P–T histories emphasizes the fact that a range of tectonic processes can lead to partial melting. In the exhumed High Himalayan Crystalline (HHC) complex, recent studies of anatexis and leucogranite magmatism have highlighted different P–T paths for anatexic rocks exposed at similar structural levels. Different P–T histories are documented for melts and migmatites derived from crystalline anatexis containing (both peritectic and cotectic) andalusite (Visonà et al., 2012), cordierite (Groppo et al., 2013), and sillimanite (Searle et al., 2010; Groppo et al., 2012) at similar structural levels. These petrological
observations highlight the complex polyphase melting and structural evolution of the HHG, and other orogens such as the Sevier that may have experienced syncontractional normal faulting and exhumation. Recognizing different petrogenetic histories for discrete blocks within an orogen provides the keys for unraveling the tectonic significance of deep crustal melting and metamorphism in an exhumed orogenic belt.

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SUPPLEMENTARY DATA
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