Zircon U-Pb Ages and Hf Isotopes of the Askot Klippe, Kumaun, Northwest India: Implications for Paleoproterozoic Tectonics, Basin Evolution and Associated Metallogeny of the Northern Indian Cratonic Margin

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Zircon U-Pb ages and Hf isotopes of the Askot klippe, Kumaun, northwest India: Implications for Paleoproterozoic tectonics, basin evolution and associated metallogeny of the northern Indian cratonic margin

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Abstract Throughout the Himalayan thrust belt, klippen of questionable tectonostratigraphic affinity occur atop Lesser Himalayan rocks. Integrated U-Pb ages, Hf isotopic, and whole rock trace element data establish that the Askot klippe, in northwest India, is composed of Paleoproterozoic lower Lesser Himalayan rocks, not Greater Himalayan rocks, as previously interpreted. The Askot klippe consists of 1857 ± 19 Ma granite-granodiorite gneiss, coeval 1878 ± 19 Ma felsic volcanic rock, and circa 1800 Ma Berinag quartzite, representing a small vestige of a Paleoproterozoic continental arc, formed on northern margin of the north Indian cratonic block. Detrital zircon from Berinag quartzite shows εHf1850 Ma values between −9.6 and −1.1 (an average of −4.5) and overlaps with εHf1850 Ma values of the Askot klippe granite-granodiorite gneiss (−5.5 to −1.2, with an average of −2.7) and other Paleoproterozoic-arc-related Lesser Himalayan granite gneisses (−4.8 to −2.2, with an average of −4.0). These overlapping data suggest a proximal arc source for the metasedimentary rocks. Subchondritic εHf1850 Ma values (−5.5 to −1.2) of granite-granodiorite gneiss indicate existence of a preexisting older crust that underwent crustal reworking at circa 1850 Ma. A wide range of εHf1850 Ma values in detrital zircon (−15.0 to −1.1) suggests that a heterogeneous crustal source supplied detritus to the northern margin of India. These data, as well as the presence of a volcanogenic massive sulphide deposit within the Askot klippe, are consistent with a circa 1800 Ma intra-arc extensional environment.

1. Introduction

Two geochemically distinct tectonostratigraphic units—the Neoproterozoic-Ordovician Greater Himalaya and lowermost part of the Paleoproterozoic lower Lesser Himalaya—form the high-grade metamorphic core of the Himalayan orogen in the Kumaun-Garhwal region (Uttarakhand) of northwestern India (Figure 1) [Valdiya, 1980; Srivastava and Mitra, 1994; Célérier et al., 2009; Spencer et al., 2012; Kohn, 2014]. However, similar lithologies, metamorphism, and ductile deformation make it difficult to differentiate these two similar looking rock units [Valdiya, 1980; Célérier et al., 2009; Rawat and Sharma, 2011; Patel et al., 2007, 2011, 2015] and to interpret the structural significance of klippen that carries these rocks, including the Almora, Bajthik, and Lansdowne klippen (Figure 1). These klippen occur as metamorphic rock outliers within the Lesser Himalayan tectonostratigraphic zone, and identifying their proper tectonostratigraphic affinity is crucial for structural reconstructions of the orogen in this region [Valdiya, 1980; Célérier et al., 2009; Rawat and Sharma, 2011; Patel et al., 2007, 2015]. Physical similarities in lithologic appearance necessitate use of geochemical discrimination methods, such as U-Pb zircon ages or Nd isotopic data, that can separate the Neoproterozoic-Ordovician rocks of the Greater Himalaya from the Paleoproterozoic rocks of the lower Lesser Himalaya.

The largest klippe in the region—the Almora klippe—is commonly interpreted as the southern extension of Paleoproterozoic Munsiari Formation equivalent rocks [Valdiya, 1980; Srivastava and Mitra, 1994; Célérier et al., 2009; Rawat and Sharma, 2011; Patel et al., 2015]. However, the klippe instead contains widespread Neoproterozoic (circa 900 Ma) metasedimentary and Ordovician intrusive rocks and must instead be composed of Greater Himalayan rocks, carried southward by the Main Central thrust (regionally called the Vaikrita thrust) [Mandal et al., 2015a]. This previous work on the Almora klippe raises the question of whether any of the other klippen share affinities with Munsiari Formation rocks or are instead all Greater Himalayan rocks. The occurrence of Munsiari-equivalent rocks in these other klippen would be significant for two...
reasons. First, they would provide constraints on Himalayan structural evolution. Second, their exposure over wide areas offer the possibility of finding rocks that better illuminate the otherwise obscure origins and early evolutionary history of the northern margin of the north Indian cratonic block [Bhowmik et al., 2012]. The NIB originally formed part of Earth’s ancient Paleo-Mesoproterozoic supercontinent, “Columbia,” [Rogers and Santosh, 2002] and the tectonic environment of the Paleoproterozoic margin of the NIB (i.e., origins of Munsiari Formation rocks and their stratigraphic equivalents) plays a key role in its placement within “Columbia”. This Paleoproterozoic margin has been variously interpreted as a passive margin [Brookfield, 1993; Upreti, 1999], a rift or plume, [Bhat et al., 1998; Ahmad et al., 1999; Ghosh et al., 2012; Sakai et al., 2013], or a continental arc [Kohn et al., 2010].

In this study, we focus on the Askot klippe, a crystalline klippe exposed between the Almora klippe to the south and the Munsiari thrust sheet to the north, in the Kumaun region of northwestern India near the western border of Nepal (Figure 1). We characterize the structure, stratigraphy, and magmatic-thermal event(s) of units within the Askot klippe through field mapping, zircon U-Pb and Hf analysis, and whole rock trace element geochemical analysis. We first establish that the Askot klippe is indeed composed of lower Lesser Himalayan affinity rocks, not of Greater Himalayan rocks. Then, we integrate our geochronologic and isotopic data with published data to: (1) refine Himalayan structural evolution in the region, (2) determine the origin and stratigraphic development history of the Lesser Himalayan Paleoproterozoic rocks, (3) compare the Paleoproterozoic evolution of the NIB’s northern margin with the western margin, the...
Delhi-Aravalli orogenic belt [Kaur et al., 2011, 2013; McKenzie et al., 2013], and (4) propose a tectonic and crustal evolution model of the northern Indian margin during the Paleoproterozoic in the context of similar data from the Aravalli-Delhi belt. The Askot klippe hosts a volcanogenic massive sulphide (VMS) deposit—the largest such deposit in the northwest Himalaya and within Lesser Himalayan rocks. VMS deposits generally form in extensional tectonic settings including back-arc/intra-arc basins, margins of volcanic arcs, and mid-ocean ridges [Galley et al., 2007; Huston et al., 2010]. The ages of the deposit and its host rocks provide key constraints on Paleoproterozoic magmatism and associated ore deposit style along the northern margin of the NIB and provide unique insights into the early tectonic history of this region.

2. Regional Geologic Setting and Stratigraphy

Following early workers [Heim and Gansser, 1939; Gansser, 1964], the Himalayan thrust belt is divided into four major tectonostratigraphic zones: from north to south, these are the Neoproterozoic through Eocene strata of the Tethyan Himalaya (TH), Neoproterozoic through Ordovician high-grade metamorphic rocks of the Greater Himalaya (GH), Paleoproterozoic to Paleozoic strata of the Lesser Himalaya (LH), and Cenozoic foreland basin strata of the Subhimalaya (SH). The South Tibetan Detachment system (STDS) separates TH to the north from GH to the south [Burchfiel et al., 1992]. The Main Central thrust (MCT or Vaikrita thrust) separates GH from LH [Heim and Gansser, 1939; Gansser, 1964; Valdiya, 1980; Srivastava and Mitra, 1994; Mukherjee, 2013]. The Ramgarh-Munsiari thrust (RMT) and the LH duplex (LHD) are intra-LH thrusts [Khanal and Robinson, 2013; Robinson and Pearson, 2013; Robinson and Martin, 2014; Mukherjee, 2015]. The Main Boundary thrust (MBT) separates LH from SH to the south. The Main Frontal thrust (MFT) separates SH from the Indo-Gangetic plain (Figure 1). All faults sole into the Main Himalayan thrust, the basal décollement, at depth.

Much of our research emphasizes LH rocks, so some outline of regional stratigraphy is warranted. Unfortunately, a long history of research in the area coupled with along- and across-strike differences in LH lithologies and metamorphic grade have led to a profusion of names. Here we provide only the most basic descriptions; further discussion is presented in Text S1 in the supporting information (see also Figure 2).
In northwest India, the LH comprises an 8–13 km thick succession of Paleoproterozoic through Cambrian clastic and carbonate metasedimentary rocks and Paleoproterozoic igneous rocks (Figure 2); here we divide the LH chronologically into upper LH (≤ ~1.0 Ga) versus lower LH (≥ ~ 1.6 Ga) [Hughes et al., 2005; Richards et al., 2005; McKenzie et al., 2011]. From stratigraphically lowest to highest, the lower LH consists of the Berinag, Damtha, and Tejam Groups. The Berinag Group contains disparate siliciclastic, volcanic, and intrusive rocks

![Geologic map of the Askot klippe. Geology draped over terrain model. The map key contains a generalized stratigraphy of the Askot klippe. Damtha Gp* of rocks are not exposed within the klippe. A-A’ is the cross-section line for Figure 4. Abbreviations: BT: Berinag thrust and SCT: South Chiplakot thrust. The solid white polygon demarcates the Askot volcanogenic massive sulphide (VMS) deposit boundary.](image)
Metamorphic grade ranges from greenschist- to upper amphibolite-facies. Age constraints regionally and from laterally correlated rocks include an interbedded basalt (1800 ± 13 Ma), numerous orthogneisses (1.80–1.85 Ga), and detrital zircon (1.88–1.96 Ga) [Miller et al., 2000; Célérier et al., 2009; Kohn et al., 2010; Long et al., 2011; Webb et al., 2011]. The greenschist-facies Damtha Group unconformably overlies the Berinag Formation [Célérier et al., 2009] and consists of siliciclastic units of the Chakrata (≤1.8 Ga) and Rautgara (≤1.6 Ga) Formations [McKenzie et al., 2011; Mandal et al., 2015a]. The greenschist-facies Tejam Group conformably overlies the Damtha Group and spans the lower to upper LH chronologic boundary. Basal Deoban/Gangolihat Dolomites are ≤1.6 Ga, whereas the slate and phyllite of the overlying Mandhali Formation are ≤~1.0 Ga [McKenzie et al., 2011].

3. Geology of the Askot Klippe

The Askot klippe occupies an asymmetrical synform in the northern part of the thrust belt (Figures 1 and 3). We combine existing maps [Valdiya, 1980; Srivastava and Mitra, 1994; Célérier et al., 2009] with new field and chronologic data to construct a cross section (Figures 3 and 4) and reassess stratigraphic and structural configurations. The core of the klippe exposes the Askot granite [Valdiya, 1980], a metaluminous, alkali granite-granodiorite gneiss (Figure 5a) [Rao and Sharma, 2009] with an L-S tectonite fabric, and a whole-rock Rb-Sr isochron age of 1983 ± 80 Ma [Bhanot et al., 1980]. The L-S fabric is well developed toward the margin of the alkalic Askot granite but poorly developed in the core. Underneath the Askot granite, the klippe contains hanging wall rocks of the Berinag thrust, structurally atop younger carbonaceous phyllite of the Mandhali Formation (Figures 4 and 5b; locally known as Tejam phyllite) [Valdiya, 1980]. The Berinag thrust sheet rocks include quartzite, schist, amphibolite, and biotite-augen gneiss (Figure 3), while the schistose rocks include chlorite-actinolite-tremolite-epidote-quartz schist, biotite-muscovite-epidote-quartz schist, and garnetiferous biotite-muscovite-chlorite-sericite-quartz schist. As there is no available study that focuses on the metamorphism of the Askot klippe, metamorphic rocks of this klippe are often correlated with greenschist- to amphibolite-grade Almora klippe rocks [Joshi and Tiwari, 2009] to the south. The primary contact between Askot granite and underlying Berinag thrust hanging wall rocks is obliterated due to Himalayan deformation. The southern contact between the granite-granodiorite augen gneiss and the underlying mica schist is sheared, with a top-to-the-north sense of shear (Figure 5c). The southern limb of the klippe dips moderately (35°) north, whereas the northern limb is overturned to steeply south dipping. Lithologically, the lower part of the klippe consists of clean cross-bedded quartzite (Figure 5d) with subordinate schist, whereas the upper part is dominated by light colored, sericitic schist (Figure 5e), yielding an overall upward fining sequence. Both the quartzite and schist are intruded by mafic sills (Figure 5f), which are strongly foliated in places to form chlorite-biotite schist. Mafic (amphibolite) sills are more concentrated upsection. In the upper part of the sequence, a biotite augen gneiss contains abundant “quartz eyes” in both hand specimen and thin section and also exhibits some compositional layering (Figures 6a–6c). This augen gneiss is intercalated with metapelites (mica-schist) and minor quartzite/sericite quartzite. Early mesoscopic, isoclinal folds with subhorizontal axes develop within incompetent mica-schist, while the late, large-scale synformal geometry develops within
the more competent quartzite and schistose quartzite, due to the growth of the Lesser Himalayan duplex since mid-Miocene time [Robinson and McQuarrie, 2012; Mandal et al., 2015b]. The bedding in the quartzite and schistose quartzite are roughly parallel to the mylonitic foliation of granite-granodiorite gneiss in the central part, suggesting that both fabrics developed together during the Cenozoic Himalayan deformation.

The Askot polymetallic (Cu-Zn-Pb ± Ag ± Au) sulphide deposit is located in the easternmost edge of the Askot klippe (Figure 3). This Zn-rich deposit is the largest known sulphide deposit in the northwest Himalaya with a reserve of 1.86 million tons of copper, zinc, lead, silver, and gold [Boswell, 2006]. Sulphide mineralization is hosted within the chloride-biotite-muscovite schist, tuffaceous (sericitic) schist, and biotite-augen gneiss, which are intensively altered (Figure 7). Chalcopyrite, galena, and sphalerite are major ore minerals, with minor amounts of pyrrhotite and arsenopyrite [Boswell, 2006]. Mineralization forms massive, lenticular sulphide lenses whose form likely reflects Himalayan deformation and metamorphism, with an average thickness of 2.5 m and a strike length of 645 m [Boswell, 2006].

Figure 5. Field photos of the Askot klippe rocks. (a) alkali granite-granodiorite gneiss from the central part of the klippe with ductilely deformed potassium feldspar porphyroclast, 20 mm diameter coin for scale. (b) Berinag thrust, which placed Berinag Formation quartzite on top of the Mandhali Formation phyllite to the south of the klippe. D. Robinson for scale. (c) Sheared contact of foliated granite-granodiorite (GG) augen gneiss and muscovite-biotite (MS) schist, 32 cm long hammer for scale, aligned parallel to the foliation plate. (d) Cross-bedded very coarse grained, clean Berinag quartzite from the southern part of the klippe; 15 cm ruler for scale. (e) Isoclinal folded chlorite-biotite schist, the host rocks for mineralization, from the underground mine; 15 cm ruler for scale. (f) Interlayered thin-bedded quartzite (Qtz) and amphibolite (Amp) sill from the northern part of the klippe; 18 cm hammer for scale.
Three samples were collected from representative Askot klippen units, one from the structurally high granite-granodiorite gneiss (SM11-028), one from the structurally intermediate biotite-augen gneiss (SM10-026) that hosts the Askot VMS deposit, and one from the structurally lower quartzite (SM10-021; Table 1). We collected 1–2 kg rock samples and separated zircon grains by using a jaw crushe, disk mill, water table, magnetic separator, and heavy liquids. A split of the zircon grains was mounted in epoxy together with Sri Lanka and R33 zircon standards [Gehrels et al., 2008]. These mounts were polished to expose zircon grains and imaged using backscattered electrons and cathodoluminescence (Figure 8) at the University of Arizona. All three samples were analyzed for U-Pb ages, and two were also analyzed for Lu-Hf (SM11-028 and SM10-021).

Three samples previously analyzed for U-Pb ages [Mandal et al., 2015a] were additionally analyzed for Lu-Hf isotopes: samples SM11-004 (Nagthat quartzite, ≤ ~1860 Ma), SM11-022 (Rautgara quartzite, ≤ ~1780 Ma), and SM11-009 (granite augen-gneiss; 1866 ± 19 Ma, including ±1% standardization error). Locations of these latter samples are shown in Figure 1.

Additional analytical details are provided elsewhere (supporting information Text S2), but briefly, zircon grains were analyzed for U-Pb ages and Lu-Hf isotopes (Table 1) via laser ablation multicollector inductively coupled plasma mass spectrometry at the University of Arizona Laserchron Facility following methods described in Gehrels et al. [2006, 2008, 2011; supporting information Table S1]. Instrumentation included a 193 nm ArF excimer laser (Photon Machines Analyte G2) and Nu HR Inductively Coupled Plasma Mass Spectrometry. Spot sizes were 30 μm for U-Pb data and positioned relative to cathodoluminescence images to ensure that the ablation pits did not cross multiple age domains or inclusions. Data were standardized to Sri Lanka zircon (563 ± 3.2 Ma; 2σ), with a standardization error of ~ ±1% (circa 15–20 Ma). Ages were based on 207Pb/206Pb, omitting highly discordant and compositionally anomalous data. Age uncertainties are reported both for precision and including a 1% standardization error. Lu-Hf spots were 40 μm, centered on previous U-Pb analyses, and positioned relative to cathodoluminescence images to ensure that the ablation pits did not cross multiple age domains or inclusions. Data were standardized to 176Hf/177Hf ratios at the time of crystallization from measurement of present-day 176Hf/177Hf and 176Lu/177Hf ratios and the age of the corresponding spot, assuming a 176Lu decay constant of 1.867 × 10⁻¹¹ a⁻¹ [Scherer et al., 2001; Söderlund et al., 2004]. Uncertainty in the 176Hf/177Hf ratio was approximately 0.00007 (2σ) or ~ ±2.5 epsilon units. Four representative whole rock samples of granitoids (SM11-009, SM10-026, and SM11-028) and a metapelite (SM11-027) of ~30 grams each (Table 1) were analyzed to obtain
rare earth element (REE) and trace element concentrations using standard methods at the University of Alabama [Stowell et al., 2010, Table 2]. The precision of REE and trace element data is ≤5%, except La and Hf (≤10%). Unless otherwise specified, all errors are reported at 2σ confidence.

5. Results
5.1. U-Pb Zircon Geochronology
Zircon grains from the granite-granodiorite gneiss (SM11-028) are prismatic (Figure 8) with oscillatory zoning and contain low to high U concentrations (400–1600 ppm; one grain 2744 ppm; see Table S1). U/Th ratios of zircon from this igneous sample range from 3.7 to 23.4 with an average of 9.9. Coherent analyses cluster tightly on or around Concordia with an upper intercept of 1855.9 ± 2.1 Ma and a lower intercept of 40 ± 5.3 Ma (Figure 9a). The weighted mean of 20 ages (>90% concordance) is 1856.6 ± 1.0 Ma (mean square weighted deviate (MSWD) = 3.8, Figure 9b), suggesting a crystallization age with no inherited cores. Our preferred age, including standardization errors, is 1857 ± 19 Ma, considerably older than Rb-Sr whole rock ages of 1380–1690 Ma [Rao and Sharma, 2009].

Zircon grains from the biotite augen gneiss (SM10-026) that hosts the Askot VMS deposit are prismatic with oscillatory zoning (Figure 8), confirming an igneous origin, and contain low to moderate U concentrations (190–769 ppm, one grain 1181 ppm). U/Th ratios range from 1.4 to 13.4, with an average of 3.3. Coherent analyses cluster tightly on or around Concordia with an upper intercept of 1879 ± 14 Ma and a lower intercept at the origin (Figure 10a). Two inherited cores yielded older ages (2448 ± 10 Ma and 2193 ± 9 Ma; see Table S1). The weighted mean of 18 ages (>90% concordance), excluding inherited cores, is 1878.4 ± 2.4 Ma (MSWD = 3.6, Figure 10b). Our preferred age, including standardization errors, is 1878 ± 19 Ma.

Detrital zircon from Berinag Formation quartzite (SM10-021) exhibits a well-defined peak around circa 1860 Ma, with a small peak around 2500 Ma and no grains younger than 1771 ± 2.2 Ma (86% concordance, Figure 11). The weighted mean of the three youngest zircon grains with measurement errors that overlap at the 1σ level is 1772 ± 17 Ma (MSWD = 15). Following the approach of Martin et al. [2011], we add the

Table 1. Analyzed Geochemical Sample Locality Information

<table>
<thead>
<tr>
<th>Sample No</th>
<th>δN (dd.ddddd)</th>
<th>δE (dd.ddddd)</th>
<th>Lithology</th>
<th>Analysis</th>
<th>Locality Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>SM10-021</td>
<td>29.7627</td>
<td>80.3474</td>
<td>Bedded quartzite</td>
<td>U-Pb-Hf (DZ)</td>
<td>Askot klippe—Berinag Quartzite</td>
</tr>
<tr>
<td>SM10-026</td>
<td>29.7581</td>
<td>80.3314</td>
<td>Biotite-augen gneiss</td>
<td>U-Pb-Hf (IZ), and trace elements</td>
<td>Askot klippe—Askot volcanics and/or volcanoclastic</td>
</tr>
<tr>
<td>SM11-027</td>
<td>29.7560</td>
<td>80.3350</td>
<td>Chlorite-biotite-muscovite schist</td>
<td>Trace elements</td>
<td>Askot klippe—schist</td>
</tr>
<tr>
<td>SM11-028</td>
<td>29.7607</td>
<td>80.3187</td>
<td>Granite-granodiorite (GG) gneiss</td>
<td>U-Pb-Hf (IZ) and trace elements</td>
<td>Askot klippe—granite-granodiorite gneiss</td>
</tr>
<tr>
<td>SM11-009a</td>
<td>29.2149</td>
<td>80.0544</td>
<td>Granite-granodiorite (GG) gneiss from Ragah Thrust sheet</td>
<td>U-Pb-Hf (IZ) and trace elements</td>
<td>South of the Almora klippe—Debguru porphyry</td>
</tr>
<tr>
<td>SM11-022a</td>
<td>29.5137</td>
<td>80.1289</td>
<td>Bedded muddy quartzite</td>
<td>U-Pb-Hf (DZ)</td>
<td>North of the Almora klippe—Rautgara Fm.</td>
</tr>
<tr>
<td>SM11-004a</td>
<td>29.1883</td>
<td>80.0918</td>
<td>Bedded clean quartzite with penecontemporaneous mafic sills</td>
<td>U-Pb-Hf (DZ)</td>
<td>South of Almora quartzite—Bhowali Quartzite (= Paleoproterozoic Berinag?)</td>
</tr>
</tbody>
</table>

a= U-Pb ages published in Mandal et al. [2015a]. DZ = detrital zircon and IZ = igneous zircon.

Figure 7. Underground exposure of the massive sulphide with pink orange oxidation stains in the Askot underground mine. The 18 cm hammer head for scale.
uncertainty and round up to the nearest 10 Ma to obtain circa 1800 Ma as the maximum depositional age of the Berinag Formation quartzite from the klippe. Including standardization errors yields a similar circa 1800 Ma maximum depositional age.

5.2. Zircon Hf Isotopes

$^{176}$Hf/$^{177}$Hf values of circa 1850 Ma zircon from granite-granodiorite gneiss sample SM11-028 (see Table S2) range from 0.281457 to 0.281563, with an average of 0.281525, and $\varepsilon$Hf(T) values range from $-5.5$ to $-1.2$, with an average ($n=12$) of $-2.7$. $^{176}$Hf/$^{177}$Hf(T) values of circa 1850 Ma zircon from augen gneiss sample SM11-009 (south of the Almora klippe; Figure 1) range from 0.280865 to 0.281538, with an average of 0.281403, and $\varepsilon$Hf(T) values range from $-4.8$ to $-2.2$, with an average ($n=7$) of $-4.0$. One inherited core with a crystallization age of $2193 \pm 58$ (93% concordance) yields an $\varepsilon$Hf(T) value of $-18.3$. Thus, overall igneous crystallization $^{176}$Hf/$^{177}$Hf(T) values for zircon are consistently albeit slightly subchondritic.

$^{176}$Hf/$^{177}$Hf(T) values of detrital circa 1850 Ma zircon from Berinag Formation quartzite sample SM10-021 (see Table S2) range from 0.280969 to 0.281562, with an average of 0.281360, and $\varepsilon$Hf(T) values range from $-9.6$ to $-1.1$, with an average of $-4.5$. $\varepsilon$Hf(T) values of circa 2500 Ma zircon range from $-7.3$ to $+4.9$, with an average of $-3.8$. $^{176}$Hf/$^{177}$Hf(T) values of circa 1850 Ma detrital zircon from Rautgara Formation quartzite sample SM10-022 (see Table S2) range from 0.280838 to 0.28942, with an average of 0.281462,

![Figure 8. CL images of igneous zircons from samples SM11-028 (granite-granodiorite gneiss) and SM10-026 (biotite augen gneiss). Solid circles represent position of U-Pb laser spot with $^{207}$Pb/$^{206}$Pb ages and 1σ error. Dashed circles represent position of Lu-Hf laser spot with initial $^{176}$Hf/$^{177}$Hf calculated using Pb-Pb age obtained from the same spot.](image-url)
and εHf (T=1850 Ma) values range from −15.1 to +7.6, with an average of −2.9. εHf of approximately 2500 zircon range from −12.1 to +6.1. Thus, overall detrital zircon 176Hf/177Hf(T) values scatter widely (subchondritic to superchondritic) with slightly subchondritic average values, similar to zircon from coeval igneous rocks.

5.3. Trace Element Geochemistry

Granitoids and the metapelite show relative enrichment in light REE (LREE) with (La/Yb)N values of 11.3 to 19.0 (Figure 12a and Table 2), suggesting a continental crustal source. All samples exhibit negative Eu anomalies [(Eu/Eu*)N = 0.18–0.82], indicating either residual plagioclase during fractional melting or plagioclase fractionation during crystallization. Partial melting should enrich incompatible large-ion lithophile elements (LILE), like K, Ba, Rb, Sr, and Eu, in the melt, while the compatible high field-strength elements (HFSE), like Th, U, Ce, Zr, Ti, Nb, Ti, Nb, Sr, and Zr anomalies and high LILE/HFSE ratios (Figure 12b) suggest that granitoids and source rocks of the metasedimentary rock formed in a continental arc, possibly through remelting of preexisting continental crust.

6. Discussion

6.1. Stratigraphic Affinity and Structure of the Askot Klippe

Pronounced differences in depositional ages and spatial association with circa 1850 Ma versus circa 500 Ma granitoids can discriminate Paleoproterozoic lower LH rocks from Neoproterozoic to Cambrian GH and upper LH rocks [Ahmad et al., 2000; DeCelles et al., 2000; Richards et al., 2005; Tobgay et al., 2010; McKenzie et al., 2011; Long et al., 2011; Khanal et al., 2015]. The preferred crystallization ages of 1857 ± 19 and 1878 ± 19 Ma for orthogneisses in the Askot klippe correspond well with circa 1850 Ma granite-granodiorite plutons and volcanic rocks associated with the lower LH [Kohn et al., 2010, and references therein]. In sample SM10-026, abundant quartz eyes (phenocrysts? Figure 6c) and prismatic zircon grains with distinct oscillatory zoning suggest that it may be a metamorphosed felsic...
volcanic or volcaniclastic rock. The Berinag Formation quartzite (SM10-021) yields Paleoproterozoic ages with no grains younger than 1771 Ma and a preferred maximum depositional age of circa 1800 Ma, indistinguishable from quartzite ages reported from lower LH rocks elsewhere along the Himalayan arc [DeCelles et al., 2000; Richards et al., 2005; Célérier et al., 2009; McKenzie et al., 2011; Long et al., 2011; Spencer et al., 2012; Khanal et al., 2014]. These age characteristics indicate that the Askot klippe is the southern continuation of the Paleoproterozoic Munsiari Formation (i.e., basal LH) [Valdiya, 1980], certainly not GH rocks. Thus, the Askot klippe cannot be correlated with the Almora klippe to the south, because the latter is associated with characteristic GH rocks: Cambro-Ordovician granite, Neoproterozoic metasedimentary rocks, and Nd isotopic and U-Pb zircon signatures of the GH [Mandal et al., 2015a].

The presence of 1857 ± 19 Ma granite-granodiorite gneiss (SM11-028) atop ≤1800 Ma quartzite (SM10-021) implies a thrust beneath the gneiss that projects to the north beneath granite-granodiorite gneisses of the Chiplakot klippe (Figure 4), which are also Paleoproterozoic in age [Mandal, 2014]. The biotite augen orthogneiss (SM10-026) crystallized at 1878 ± 19 Ma and structurally overlies younger circa 1800 Ma Berinag Formation quartzite. This age inversion (older rocks over younger rocks) suggests a possible structural inversion. We further infer an out-of-sequence thrust on the northern side of the klippe; the younger Mandhali Formation dips northward beneath the older Deoban Formation, and to the northwest, the inferred thrust cuts the Berinag thrust, part of the regional RMT roof thrust [Mandal et al., 2015b]. This out-of-sequence thrust cuts both the Berinag Formation and overlying thrusts that contain granite-granodiorite gneiss (Figure 4).

6.2. Origin and Stratigraphic Development of the Paleoproterozoic Lower Lesser Himalayan Rocks

Trace element tectonic discrimination diagrams (Figures 13a and 13b) show that the 1857 ± 19 Ma granite-granodiorite gneiss of the Askot klippe plots within volcanic arc or syncollisional field, similar to other Askot gneiss samples [Rao and
This rock is strikingly similar to the Debguru porphyry (SM11-009, Figure 1), a granodiorite gneiss in the Ramgarh-Munsari thrust hanging wall, south of the Almora klippe, with a crystallization age of 1852 ± 19 Ma [Mandal et al., 2015a]; (error includes calibration uncertainties). Zircon from both rocks has εHf1850 Ma values between approximately −2 and −5 and both plot within the volcanic arc or syn-collisional field in tectonic discrimination diagrams (Figures 13a and 13b). Several studies emphasize characteristic calcalkaline trace element patterns for these rocks, including LILE enrichments and HFSE depletions (e.g., Figure 13), consistent with an arc origin [see Rao and Sharma, 2009, 2011]. Following others [Rao and Sharma, 2009, 2011; Kohn et al., 2010; Kaur et al., 2013; Sen et al., 2013], we generally interpret these rocks as having formed in a circa 1850 Ma arc.

More regionally, εHf values of circa 1850 Ma zircon from our sample of Berinag Formation quartzite (−9.6 to −1.1, with an average of −4.5) overlap with the εHf value of the granite-granodiorite gneiss of the Askot klippe (SM11-028; εHf = −5.5 to −1.2; average = −2.7), the Debguru porphyry granodiorite gneiss (SM11-009; εHf = −4.8 to −2.2; average = −4.0), and the 1850–1900 Ma Wangtu orthogneiss (−0.9 to −2.8) and Rampur Formation metasedimentary rock (−1.5 to −3.7) [Richards et al., 2005] from the Sutlej Valley, approximately 500 km to the northwest, and confirm the correlation of the Askot klippe with lower LH rocks. Isotopic consistency, the presence of a tuffaceous (sericitic) schist, and similar calcalkaline trace element trends among the gneisses and metapelite that is interbedded with the quartzite, further support a proximal arc source of the Berinag Formation quartzite (Figure 12). Older zircon grains (circa 2500 Ma) from a Berinag Formation quartzite (SM11-021) have εHf values ranging from −7.3 to −2.1 (supporting information Table S2) and suggest derivation from reworked Archean NB, which experienced major crustal reworking and copious granite magmatism at circa 2500 Ma [Mondal et al., 2002].

Table 2. Whole Rock REE and Trace Element Compositions

<table>
<thead>
<tr>
<th>Analyte</th>
<th>SM11-009</th>
<th>SM10-026</th>
<th>SM11-027</th>
<th>SM11-028</th>
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<tbody>
<tr>
<td>Rb</td>
<td>121.42</td>
<td>141.89</td>
<td>211.48</td>
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<tr>
<td>Y</td>
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<td>27.10</td>
<td>37.90</td>
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<tr>
<td>Nb</td>
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<td>13.86</td>
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<tr>
<td>La</td>
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<td>46.66</td>
<td>70.09</td>
<td>17.11</td>
</tr>
<tr>
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<tr>
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<tr>
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</tr>
<tr>
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<tr>
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<tr>
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<tr>
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<tr>
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<tr>
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<tr>
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<td>0.68</td>
<td>0.82</td>
<td>0.18</td>
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<tr>
<td>(La/Yb)N</td>
<td>19.03</td>
<td>14.16</td>
<td>14.31</td>
<td>11.39</td>
</tr>
</tbody>
</table>

Figure 13. Trace element discrimination diagrams [Pearce et al., 1984], showing syncollisional/arc-related source of these samples.
Overall, we interpret that geographically widespread approximately 1850 granitoid plutons as forming the base of the Berinag, Munsiari, and Ramgarh Formations and these orthogneisses are all genetically equivalent, just currently residing on different thrust sheets—the Ramgarh-Munsiari thrust and the Berinag thrust. Paleoproterozoic granitoids were once part of the NIB’s northern margin continental arc, which is now dissected because of Cenozoic Himalayan thrusting. These gneisses probably formed as shallow plutons that intruded the base of the NIB’s northern margin Paleoproterozoic cover sequence. The profuse circa 1850 Ma felsic magmatic event links to the amalgamation of the Paleo-Mesoproterozoic supercontinent “Columbia” [Rogers and Santosh, 2002, 2009; Zhao et al., 2004; Hou et al., 2008; Kohn et al., 2010]. However, we argue that the granite-granodiorite gneiss is not a part of the NIB’s basement [Célérier et al., 2009] in a strict sense, which instead consists of tonalite-trondhjemite-granodiorite with an age of ≥ circa 2.5 Ga [Mondal et al., 2002].

6.3. Implication for Paleoproterozoic Crustal Evolution of the Northern Margin of the NIB

Zircon U-Pb data alone are not capable of distinguishing grains with similar ages derived from different sources nor do they provide any information on whether the crust is juvenile or reworked. In this study, combined U-Pb zircon ages (Figure 14) and Lu-Hf data (Figures 15a and 15b) help to establish the Paleoproterozoic evolution of the NIB’s northern and western margins (lower LH and Delhi-Ararvalli orogenic belt, respectively). Using U-Pb ages and Lu-Hf isotope compositions from Delhi-Ararvalli orogen and lower LH, Kaur et al. [2013] suggested that a contiguous circa 1850 Ma subduction system bounded the northern and western margins of the NIB. However, their work was based on one sample of the Munsiari Formation that has no pre-1900 Ma grains [Spencer et al., 2012]. A composite normalized-probability age plot of lower LH detrital zircon (Figure 14) yields a strong peak at approximately 1850, suggesting that the Berinag Formation (this study) [McKenzie et al., 2011], Munsiari Formation [Spencer et al., 2012], and a Paleoproterozoic quartzite
Mandal et al., 2015a] received detritus from proximal approximately 1850 volcanic or plutonic sources. Only the Munsiari Formation in Garhwal is devoid of any pre-1900 Ma grains, suggesting that using the Munsiari Formation alone for crustal evolution correlation studies may misrepresent age and isotopic signals. A wide range of negative $\varepsilon$Hf values from approximately 1850 detrital zircon (SM11-022, −15.0 to −1.6 and SM10-021, −9.1 to −1.1) suggests a heterogeneous crustal source. This more comprehensive Hf isotopic dataset for circa 1850 Ma lower LH rocks (Berinag/Munsiari Formation) and Aravalli quartzite (Figure 15b) supports previous inferences of heterogeneous crustal sources [Kaur et al., 2011, 2013]. Highly negative $\varepsilon$Hf values for zircon from the Munsiari Formation and circa 1850 Ma granitoid plutons (Figure 15b) suggest that preexisting crust underwent crustal reworking at circa 1850 Ma. Positive and negative $\varepsilon$Hf values of circa 2500 Ma zircon from lower LH and Aravalli rocks are consistent with shared, heterogeneous sources, including contributions of juvenile and reworked material (Figure 15b). Younger (circa 1600 Ma) zircon from the Paleoproterozoic Rautgara Formation quartzite (SM11-022) also show positive to negative $\varepsilon$Hf (17.2 to +7.5; Figure 15b), again suggesting juvenile and reworked components. Overall, circa 1850 Ma zircon is consistent with reworking of preexisting crust along both the northern and western margins of the NIB, whereas older and younger zircon suggests additional juvenile mantle contributions, possibly under a different tectonic regime.

6.4. Askot Volcanogenic Massive Sulphide Deposit

Mineral deposits help to determine tectonic setting, environmental conditions, and changes in Earth’s thermal history. Supercontinent cycles can be correlated using the distribution of ore deposits because tectonic settings of ore deposits are tied to plate margin processes [Cawood and Hawkesworth, 2013]. The Askot VMS sulphide deposit is hosted within the upper part of a supracrustal sequence (i.e., chlorite-biotite-muscovite schist, tuffaceous (sericitic) schist, and granitic orthogneiss). The trace element composition of the gneiss is consistent with formation in a volcanic arc or syncollisional tectonic setting (Figure 13). Similar REE trends and trace element ratios in granitoids, augen gneiss (felsic volcanic rocks), and metapelite (Figures 11 and 12) suggest proximal sources for Paleoproterozoic metasedimentary rock of the basal unit of the lower LH rocks. Given these lithologies and geochemical characteristics, we rule out a mid-ocean ridge as a tectonic environment for the formation of this deposit and consider other tectonic scenarios.

The biotite-augen gneiss (SM10-026), one of the mineralized host rocks, yields a preferred crystallization age of 1878 ± 19 Ma, suggesting that it was coeval/comagmatic with the Askot granite-granodiorite gneiss (1857 ± 19 Ma). This biotite augen gneiss probably has an altered volcanic and/or volcanioclastic precursor, as is
evident from (1) presence of abundant quartz eyes (Figure 6) and (2) well-developed prismatic zircon grains with oscillatory zoning (Figure 8) and igneous Th/U ratios (0.16 to 0.49, with one grain 0.07) [Hoskin and Schaltegger, 2003]. Extensive sericitic alteration of feldspars may indicate later hydrothermal activity. We infer that the age of the Askot deposit is younger than 1878 ± 19 Ma and probably synchronous with the extensional event at 1800 ± 13 Ma [Miller et al., 2000] that is marked by abundant mafic sills within Berinag Formation.

Elsewhere in the Himalaya, Paleoproterozoic deposits occur at Gorubathan (Darjeeling, West Bengal, and India) and Rangpo (east Sikkim) and are hosted within the siliciclastic-mafic dominated Paleoproterozoic basal unit of the lower LH, with model ages of 1800 Ma [Sarkar et al., 2000]. The Rangpo deposit is hosted within the Gorubathan subgroup of Paleoproterozoic Daling Formation, the lower unit of the Daling-Shumar Group [McQuarrie et al., 2008]. Host rocks of Rangpo mineralization include chlorite-quartz schist, phyllite, quartz-sericite schist, sericitic quartz schist, and quartzite [Sarkar et al., 2000], similar to the arc-derived sedimentary rock of the Askot klippe. In addition, the Rangpo and Gorubathan deposits are spatially associated with Lingtse gneiss, dated at circa 1850 Ma [Mottram et al., 2014; Bhattacharyya et al., 2015]. The Rangpo deposit is interpreted to be of sedimentary diagenetic origin; however, its strong resemblance in lithology and spatial association of circa 1850 Ma arc related granitoids as is found in the Askot klippe leads us to reinterpret the Askot, Rangpo, and Gorubathan deposits as similar kinds of volcanogenic/volcanic-hosted sulphide deposits that formed at circa 1800 Ma in an extensional setting.
6.5. Paleoproterozoic Tectonic Model of the NIB’s Northern Margin

Our preferred tectonic model begins with a well-established, circa 1850 Ma arc on 2.5 Ga rocks along the northern margin of the NIB (Figure 16a) [Kohn et al., 2010], followed by extension beginning circa 1800 Ma. This extension may have developed in response to a switch in tectonic styles (e.g., termination of subduction) or slab roll back at ~1800 Ma (Figure 16b), as indicated by 1800 ± 13 Ma mafic sills [Miller et al., 2000], which are abundant within the Berinag Formation and equivalent siliciclastic rocks along the arc. Although Kohn et al. [2010] argue for arc activity until 1780 Ma, typical minimum age uncertainties of 1–2% allow termination of felsic arc magmatism as early as 1800 Ma. Age of the extension-related Askot VMS deposit can be bracketed as circa 1800 Ma (including standardization error) when arc-related late stage plutons served as possible heat source for the VMS-forming hydrothermal system. The fining-up sequence of the Askot supra-crustal metasedimentary rock indicates the Lesser Himalayan basin deepened progressively, culminating in the deposition of the circa 1600 Ma Rautgara Formation turbidite [Mckenzie et al., 2011; Mandal et al., 2015a] toward the southern fringe of the extensional (back-arc?) basin.

7. Conclusions

Integrated U-Pb geochronologic data, Hf isotopic data, and whole rock trace element geochemical studies coupled with mapping of the Askot klippe rocks lead to following conclusions.

1. The Askot klippe consists of an 1857 ± 19 Ma granite-granodiorite gneiss, coeval 1878 ± 19 Ma felsic volcanic rock, and circa 1800 Ma quartzite and is a small vestige of a Paleoproterozoic continental arc that formed on the northern margin of the northern Indian cratonic block as a part of the Paleo-Mesoproterozoic supercontinent, “Columbia”. Overlapping 1850 Ma ages and $\epsilon$Hf1850 Ma values suggest Askot klippe rocks correlate with Munsianri Formation rocks to the north. These same rocks are exposed farther south underneath the Almora klippe as the Ramgarh-Munsianri thrust sheet.

2. The Paleoproterozoic crustal evolution of the north Indian cratonic block’s northern margin in Paleoproterozoic and Archean time is strikingly similar to that reported from Delhi-Aravalli orogen along the western margin of the northern Indian cratonic block, evident from a preponderance of circa 1850 Ma ages and similar ranges of $\epsilon$Hf values. These data suggest that a much older 2.5 Ga crust was reworked at circa 1850 Ma.

3. An extensional event initiating as early as 1800 Ma is recorded in abundant mafic sills and dikes intruding upper Berinag Formation quartzite. This event could reflect a tectonic switch from subduction to extension or continued subduction with regional back-arc extension.

4. The Neoarchean and Paleoproterozoic northern margin had very heterogeneous crustal assemblages as is evident from the large range of $\epsilon$Hf values (+7.5 to ~18.1).

References


