THE DEPTH DISTRIBUTION OF SEISMICITY AT THE NORTHERN END OF THE RWENZORI MOUNTAINS: IMPLICATIONS FOR HEAT FLOW IN THE WESTERN BRANCH OF THE EAST AFRICAN RIFT SYSTEM IN UGANDA

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Abstract
Data from a six-month deployment of seismic stations around the northern end of the Rwenzori Mountains have been used to investigate the depth extent of seismicity and its implications for heat flow in the Western Branch of the East Africa Rift System in Uganda. Previous seismicity studies of the Western Branch in Uganda show earthquake nucleation at depths greater than or equal to 40 km suggesting that heat flow from the rift is not elevated. However, heat flow elsewhere in the Western Branch (e.g., western Tanzania and Lakes Kivu and Tanganyika) is elevated, similar to the elevated heat flow in the Eastern Branch of the East African Rift System in Kenya.

To investigate further the depth extent of seismicity in the Western Branch, seismological data were collected and analysed to obtain accurate hypocenters using both standard and double difference location algorithms. Focal mechanisms were also obtained to place constraints on the local stress regime. Results show that seismicity is concentrated along the major border faults of the Rwenzori horst, suggesting an eastward dip on the Ruimi-Wasa Fault and a northwestward dip on the Toro-Bunyoro Fault. Fault motions are predominantly normal, with possible strike slip motions between the Ruimi-Wasa and Toro-Bunyoro Faults. The number of earthquakes increases with depth, peaking at 16 km and falling sharply by 22 km. Strength envelope models indicate that heat flow in the range of 54 mWm\(^{-2}\) to 66 mWm\(^{-2}\) is required to explain this result. A heat flow of 54 to 66 mWm\(^{-2}\) is similar to heat flow in other mobile belts in East Africa away from major rift valleys, and therefore heat flow around the Rwenzori Mountains does not appear to be elevated. This finding does not preclude the possibility of a deep seated (i.e., mantle) thermal anomaly beneath the rift, which may not have had sufficient time to reach the surface.

Introduction
In this study, data from a six-month deployment of seismic stations around the northern end of the Rwenzori mountains have been used to investigate the depth extent of seismicity and its implications for heat flow from the Western Branch of the East African Rift System (EARS) in Uganda. Previous seismicity studies of the Western Branch in Uganda show earthquake nucleation at depths greater than or equal to 35 km (Albaric et al., 2008; Maasha, 1975; Ochman et al., 2007), suggesting that heat flow from the rift is not elevated. The average heat flow in Proterozoic mobile belts in East Africa away from the rift valleys is 63 mWm\(^{-2}\) (Nyblade, 1997), and Nyblade and Langston (1995), using a rheological model for the crust with a brittle upper crust and ductile lower crust (Brace and Kohlstedt, 1980; Byerlee, 1968; Ranalli 1987), showed that for a heat flow of 63 mWm\(^{-2}\), brittle deformation, and therefore seismicity, can extend to crustal depths of ~30 to 35 km.

However, heat flow elsewhere along the Western Branch is elevated (~109 mWm\(^{-2}\) in western Tanzania; 73 mWm\(^{-2}\) and 67 mWm\(^{-2}\) in Lakes Kivu and Tanganyika, respectively; Nyblade et al., 1990), similar to the elevated heat flow (100 mWm\(^{-2}\)) in the Eastern Branch of the EARS in Kenya (Wheildon et al., 1994). And for a heat flow of ~100 mWm\(^{-2}\), it is unlikely that brittle deformation and therefore seismicity would extend much deeper than about 10 km.

To investigate further the depth extent of seismicity in the Western Branch, seismic data were collected and analysed to obtain accurate hypocenters from the northern part of Rwenzori Mountains. Events were located using both standard and double difference location algorithms. Focal mechanisms were also obtained to place constraints on the local stress regime. Results show that seismicity, which is concentrated along the major border faults of the Rwenzori horst, extends to lower crustal depths. This result is analysed using strength envelope models to place new constraints on heat flow from the Western Branch in Uganda.
rift and represents the incipient stages of continental break up (Fairhead and Stuart, 1982). The Eastern Branch of the EARS extends from the Afar Depression in Ethiopia, through Kenya and into Tanzania (Gregory, 1921; Sikes, 1936) (Figure 1). The Western Branch of the EARS extends from southern Sudan through western Uganda, Rwanda, Burundi and Tanzania along the boundary with the Democratic Republic of Congo (Figure 1).

The EARS formed within the Precambrian basement of East Africa, which consists of the Archean Tanzania craton in the centre of the region surrounded by a number of Proterozoic mobile belts (Figure 1). The formation of the two branches of the EARS may have begun at different times but both branches experienced significant tectonic activity in the late Cenozoic (Chorowicz, 2005; Ebinger, 1989; Maasha and Molnar, 1972).

The main geological feature within the study area is the Rwenzori Mountains, a large horst block about 120 km long and 65 km wide (Figures 2 and 3). The mountain range consists of late Archean and early
Figure 2. Geology of northern Rwenzori Mountains adapted from an open report from the Department of Geological Survey and Mines, Uganda.
Proterozoic basement rock uplifted 5,000 m above sea level (Tanner, 1973). The western side of the mountain range is bounded by the Bwamba Fault, a steeply dipping normal fault (Ebinger, 1989). The eastern side is more gradual, except for the north-northeast trending Ruimi-Wasa and Toro-Bunyoro Faults, which define the narrow northern nose of the mountain range (McConnel, 1972) (Figures 2 and 3). The Pleistocene-Holocene Kicwamba volcanic field lies just east of the Ruimi-Wasa Fault. Pleistocene-Holocene sediments, locally known as the Kaiso-Kisegi beds, occupy the rift valley floor in the Semuliki basin, with thicknesses reaching 4000 to 6000 metres (Figures 2 and 3).

A previous seismicity study in the region was conducted by Maasha (1975), which employed four short period seismographs deployed at 28 different sites throughout the Rwenzori Mountains. The sites occupied at any one time formed a quadrangle. Distances between adjacent stations ranged from 5 to 20 km, and the stations were run for 4 to 20 days at each site. Fault plane solutions from this study indicate that the Rwenzori Mountains are bounded by normal faults (Maasha, 1975). The seismicity recorded was mainly associated with the Ruimi-Wasa Fault. Events located beneath the mountain range extended to depths of 25 to 40 km.

Seven moderate earthquakes in the region have been studied in detail using teleseismic data (Shudofsky, 1985; Shudofsky et al., 1987; Upcott et al., 1996). Earthquake focal mechanisms show that the rift faults bordering the Semuliki basin and the southern end of the Lake Albert rift valley are oriented nearly north-south or northeast-southwest. An exception to this general strike direction is the north-northwest to northwest striking Kisomoro Fault, along which a M 6.2 earthquake occurred in February 1994. Thus, the region appears to be under a state of east-west or northwest-southeast extension.
**Data acquisition**

The Department of Geological Survey and Mines, Uganda, in collaboration with the Federal Institute of Geosciences and Natural Resources (BGR), Germany, conducted a passive seismic survey around the Buranga geothermal area between February and August 2006. The purpose of the seismic survey was to collect micro earthquake data to aid in geothermal prospecting. Although the stations were in operation for a period of seven months, data for only four months (March to June 2006) had reliable timing and could be used in this study.

Figure 3 shows the location of the seismic stations, which were arranged in a local network covering less than 60 square kilometers. The stations were equipped with three component short period L4-3D Mark seismometers. The data was recorded at a rate of 250 samples per second using 24-bit RefTek data loggers configured to operate in trigger mode. Global Positioning System clocks were used for timing.

**Data analysis**

The data collected were analysed to obtain event hypocenters and focal mechanisms. A standard event location programme was used to initially locate the earthquakes, and then a double difference algorithm was used to obtain relative event locations. Focal mechanisms were obtained using first motion polarities.

**Phase picking**

Pg arrival times were picked for all events. Sg was not clearly seen on most seismograms and was therefore picked for only a subset of events (74 total). A 0.5 to 5 Hz band pass filter was applied to the data prior to picking arrival times, which were picked to within 0.1 sec. Events with a minimum of four good quality picks were selected for further analysis. Figure 4 shows samples of the filtered seismograms illustrating the quality of P arrival picks.

**Initial Hypocentral locations**

Arrival times were used with the HYPOELLIPSE program (Lahr 1998) to obtain initial hypocentral locations. The velocity model used for event locations is given in Table 1.

![Sample seismogram of an earthquake recorded by the network. The seismograms are vertical velocity records filtered between 0.5 to 5 Hz.](image)

**Table 1.** Velocity model used for event locations from Langston et al. (2002).

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>P Vel (km/s)</th>
<th>S Vel (km/s)</th>
<th>Density (kg/m³)</th>
</tr>
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<tr>
<td>0 - 10</td>
<td>5.84</td>
<td>3.38</td>
<td>2.35</td>
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<td>10 - 20</td>
<td>6.26</td>
<td>3.62</td>
<td>2.5</td>
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<tr>
<td>20 - 30</td>
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<td>30 - 37</td>
<td>7.09</td>
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<td>2.84</td>
</tr>
<tr>
<td>37 - 200</td>
<td>8.28</td>
<td>4.74</td>
<td>3.51</td>
</tr>
</tbody>
</table>
locations. In this programme, hypocenters are found by minimising the root-mean-square (RMS) of the travel-time residuals. Hypocenters for a total of 574 earthquakes were obtained. The velocity model used in the analysis is from Langston et al. (2002) (Table 1), which is an average model for Precambrian terrains across eastern Africa. That model was used because a local crustal model is not available for the study area.

For epicenters located within a radius of 40 km from the centre of the network, uncertainties in epicentral locations, given by the 68% error ellipse, are ± 1 km. Uncertainties of ± 3 km are associated with epicenters more than 40 km from the centre of the network. Uncertainties in source depth are ± 1 km within 40 km of the centre of the network and increase to ± 5 km for distances more than 40 km from the centre of the network.

To further assess the variation of uncertainties with respect to the velocity model, the velocities were perturbed by ± 5% and then the events with Pg and Sg picks were relocated. After relocation it was found that the epicenters shifted by no more than 0.5 km.

**Relative event locations**

Relative event locations were obtained by using the double-difference algorithm (HYPODD) of Waldehauser (2001). The advantage of the double-difference algorithm over standard location algorithms is that when the difference between the hypocentral separations of two earthquakes is small compared to the event-station distance, then the ray paths between the source region and a common station are similar along the entire path (Walderhauser and Ellsworth, 2000). The double difference algorithm minimises residual travel time differences for a pair of earthquakes at a single station. To find the optimal correlation lengths, event-station separation distances of 20, 30, 40, and 50 km were tested along with event-event distances of 3, 5, 7, 10, and 15 km. Using correlation lengths of 40 km for event-
station pairs and 5 km for event-event pairs gave the tightest clustering of events with the fewest number of rejected events. A total of 107 events were rejected using these distances.

**Focal mechanisms**

Focal mechanisms for 69 events were obtained using P motion polarities and the FOCMEC programme (Snoke 2003). In this programme, an earthquake is modeled as a point source and a grid search is performed for all combinations of strike, dip, and rake to find all sets of the nodal planes that match the polarities. An increment of 5° was used in the grid search with a zero tolerance error. Of the 69 mechanisms computed, only nine were well constrained, giving only one set of nodal planes that matched the P polarities.

**Results**

The distribution of epicenters from HYPOELLIPSE is shown in Figure 5. The seismicity pattern correlates with the Ruimi-Wasa and Toro-Bunyoro Faults on the eastern side of the Rwenzori Mountains (Figure 3). The greatest concentration of events is found along the Ruimi-Wasa Fault south of station BUTU. The seismicity to the west of the Rwenzori Mountains falls within the Semuliki Basin rather than along the Bwamba Fault. A cluster of events is seen north of the Buranga hot spring at the northern end of the Bwamba Fault.

A plot of frequency versus depth for all earthquakes located using just P arrival times is shown in Figure 6. The number of earthquakes increases with depth, peaking at a depth of 16 km and falling off sharply by 22 km. 90% of the hypocenters are shallower than 22 km depth.

To evaluate the accuracy of the hypocenters, 74 events along the Ruimi-Wasa Fault with good quality waveforms on which S arrivals could be easily picked were relocated using both P and S arrival times and the HYPOELLIPSE code. This set of events was also relocated using P arrival times only and the HYPODD code. Maps showing epicentral locations using just P arrival times and HYPOELLIPSE, P and S arrival times and HYPOELLIPSE, and P arrival times and HYPODD are shown in Figures 7, 8 and 9, respectively, together with plots of focal depth versus frequency and cross sections showing earthquake depth along the strike of the fault.

A comparison of these figures illustrates that while the locations and depths of individual events can shift by a few kilometers, the overall pattern of seismicity (epicenters and depth) does not change significantly. Thus, the distribution of focal depths shown in Figure 6 is fairly well resolved.

Figure 10 shows the epicenter locations after relocating all the events using HYPODD. The seismicity now clusters more tightly along the Ruimi-Wasa and Toro-Bunyoro Faults. Most of the events in the Semuliki Basin were rejected by the HYPODD algorithm, except for a swarm of events just north of the Buranga hot springs.

Cross-sections through the southern and northern parts of the Ruimi-Wasa Fault and through the Toro-Bunyoro Fault are shown in Figure 11. For the southern part of the Ruimi-Wasa Fault, hypocenters shift to the east with depth, suggesting an east-south-eastward dip at an angle of about 40°. A similar east-south-eastward shift of hypocenters is found for the northern segment of the fault. For the Toro-Bunyoro Fault, the hypocenters are aligned almost vertically. None of the seismicity aligns along a plane.
Figure 7. Plot of 74 events along the Ruimi-Wasa Fault located using only P arrival times and HYPOELLIPSE. (a) Map showing epicenter locations. (b) Plot of focal depth versus number of events. (c) Cross-section showing earthquake depths along the strike of the fault. The location of the cross-section is shown by the black line in (a).
Figure 8. Plot of 74 events along the Raimi-Wasa Fault located using P and S arrival times and HYPOELLIPSE. (a) Map showing epicenter locations. (b) Plot of focal depth versus number of events. (c) Cross-section showing earthquake depths along the strike of the fault. The location of the cross-section is shown by the black line in (a).
Figure 9. Plot of 74 events along the Ruimi-Wasa Fault located using only P arrival times and HYPODD. (a) Map showing epicenter locations. (b) Plot of focal depth versus number of events. (c) Cross-section showing earthquake depths along the strike of the fault. The location of the cross-section is shown by the black line in (a).
Figure 12 shows the nine well-constrained focal mechanisms. The nodal planes shown were obtained by averaging the full range of solutions obtained from the grid search (Table 2). Mechanisms one and two are on the very northern end of the Bwamba Fault, and mechanisms six, seven, eight and nine are on the Ruimi-Wasa Fault. These mechanisms all show normal faulting with some oblique slip motion and extension in a general east-west direction. Mechanisms three, four and five show strike-slip motion. The magnitudes of earthquakes have not been determined.

Discussion
Following from the work of Nyblade and Langston (1995) for East Africa, we use the depth extent of seismicity in the study area to place an upper bound on heat flow by using strength envelope calculations to estimate the depth of the brittle-ductile transition. At shallow crustal depths, the deformational behavior of rock is dominated by “brittle” elastic frictional failure and depends on pore pressure. At greater crustal depth, ductile deformation occurs, which is dependent on rock type, strain rate and temperature (Brace and Kohlstedt, 1980; Byerlee, 1968; Ranalli 1987). In general, earthquakes are believed to occur in the “brittle” elastic frictional zone, while in the ductile zone, deformation occurs by aseismic creep.

The deformation laws used to construct a strength envelope involve two equations. The first equation, for frictional shear failure, is

\[
\sigma_r - \sigma_s = \beta \rho g z (1-\lambda)
\]

where \(\sigma_r - \sigma_s\) is the critical stress difference, \(\beta\) is a fault parameter, \(\rho\) is density, \(z\) is depth, \(g\) is the acceleration due to gravity and \(\lambda\) is the ratio of the pore fluid pressure to the overburden pressure.
Figure 11. Map and cross-sections showing epicenters (black circles) for three clusters (a1) along the southern part of Ruimi-Wasa Fault, (a2) along the middle part of Ruimi-Wasa Fault, (a3) along the Toro-Bunyoro Fault. The locations of the cross sections are shown by the grey lines.
The second equation, for failure by power-law creep, is

\[ \sigma_1 - \sigma_2 = (\dot{\varepsilon}/B)^{1/n} \exp(E/nRT) \]  

(2)

where \( \dot{\varepsilon} \) is strain rate, \( T \) is absolute temperature, \( R \) is the gas constant, and \( B \), \( n \) and \( E \) are material properties.

In this model, if the critical stress for frictional failure is less than that for ductile creep, then brittle failure will occur; otherwise deformation will occur by ductile.

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Table 2. Focal mechanism shown in Figure 12.

<table>
<thead>
<tr>
<th>Event origin time</th>
<th>Location</th>
<th>Depth</th>
<th>Strike</th>
<th>Dip</th>
<th>Rake</th>
</tr>
</thead>
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<tr>
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<td>Latitude (degree)</td>
<td>Longitude (degree)</td>
<td>(degree)</td>
<td>(degree)</td>
<td>(degree)</td>
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<tr>
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<td>55</td>
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<tr>
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<td>30.2199</td>
<td>15.2</td>
<td>281</td>
<td>37</td>
</tr>
<tr>
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<td>0.8242</td>
<td>30.2591</td>
<td>15.6</td>
<td>289</td>
<td>51</td>
</tr>
<tr>
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<td>30.2132</td>
<td>18.3</td>
<td>6</td>
<td>67</td>
</tr>
<tr>
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<td>30.1461</td>
<td>4.0</td>
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</tr>
<tr>
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<td>30.2047</td>
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<td>30.1364</td>
<td>6.0</td>
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<td>60</td>
</tr>
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<td>9 20060521 19 39:49.1</td>
<td>0.5751</td>
<td>30.1387</td>
<td>5.9</td>
<td>27</td>
<td>60</td>
</tr>
</tbody>
</table>

---

Figure 12. Well constrained focal mechanisms. Numbers on focal mechanisms correlate with event numbers in Table 2.
creep. The brittle-ductile transition is the depth at which the two equations are equal. Estimates of crustal temperatures are required for solving equation (2). With constraints on surface heat flow, crustal heat production and the thermal conductivity of crustal rocks, geotherms can be calculated. However, because there are no heat flow estimates or heat production measurements for the Western Branch in Uganda, we simply assume a linear geothermal gradient and proceed with the calculation. We also assume a quartz diorite lithology for the crust and use the rheological parameters from Hansen and Carter (1982) and Shelton and Tullis (1981); \( B = 1.3 \times 10^{-3} \), \( n = 2.4 \) and \( E = 219 \). Other parameters used include a strain rate of \( 10^{-15} \text{s}^{-1} \), a crustal density of \( 2800 \text{kgm}^{-3} \), a \( \beta \) value of 0.75, which is appropriate for normal faulting, and \( \lambda = 0.36 \) (i.e., pore pressure equal to hydrostatic pressure). A strain rate of \( 10^{-15} \text{s}^{-1} \) is constrained by estimates of crustal extension in the East African rift valleys, and away from the rifts the strain rates may be lower than that (Nyblade and Langston 1995).

Figure 13 shows strength envelopes for a range of different geothermal gradients that place the brittle-ductile transition at depths of 16 to 22 km, which is...
roughly consistent with the depth at which seismicity in the study area decreases dramatically. If we assume a thermal conductivity of 3 W/mK for upper crustal rock, then from Fourier’s law of conductive heat flow, \( Q = \kappa \frac{dT}{dz} \), heat flow can be estimated, where \( Q \) is the heat flow, \( \kappa \) is the thermal conductivity and \( \frac{dT}{dz} \) is the assumed linear geothermal gradient. A range of heat flow between 54 and 66 mWm\(^{-2}\) is indicated by the depth at which the brittle ductile transition corresponds to the depth at which seismicity falls off significantly (Figure 13). A thermal conductivity of 3 W m\(^{-1}\)K\(^{-1}\) is a reasonable average value for felsic rocks typically found in the upper crust (Nyblade and Langston, 1995). In comparison, the strength envelope for a high heat flow of ~100 mWm\(^{-2}\) gives a brittle-ductile transition that is too shallow (~10 km depth) to explain the observed depth extent of seismicity (Figure 13).

Strength envelopes using a quartz rheology are shown in Figure 14, illustrating that heat flow of 54 to 66 mWm\(^{-2}\) is an upper bound. If a quartz rheology is used, then much lower heat flow would be needed to explain the observed depth extent of seismicity.

Is heat flow between 54 and 66 mWm\(^{-2}\) elevated, as one might expect to find in a rift valley? As mentioned in the introduction, Nyblade (1997) reported a mean heat flow of 63 mWm\(^{-2}\) for all Proterozoic mobile belts in East Africa away from the main rift valleys. The heat flow estimated here for the study areas is thus similar to heat flow from other Proterozoic mobile belts and does not appear to be anomalously high. A heat flow of 54 to 66 mWm\(^{-2}\) is also consistent with average heat flow from Proterozoic terrains globally (Nyblade and Pollack, 1995). In contrast, a heat flow of 54 to 66 mWm\(^{-2}\) is somewhat lower than the heat flow in Lakes Tanganyika (73 mWm\(^{-2}\)) and Kivu (67 mWm\(^{-2}\)) (Nyblade et al., 1990), and considerably lower than a heat flow of 109 mWm\(^{-2}\) reported in the Western Branch in Tanzania (Nyblade et al., 1990) and a heat flow of 100 mWm\(^{-2}\) reported in the Eastern Branch of the EARS in Kenya (Wheeldon et al., 1994). The finding that heat flow is not elevated in the study area, however, does not preclude the possibility of a deep seated (mantle) thermal anomaly beneath the Western Branch in Uganda that has not had sufficient time to reach the upper crust.

Summary

In summary, seismicity at the northern end of the Rwenzori Mountains correlates well with the major border faults, suggesting an eastward dip on the Ruimi-Wasa Fault and a northwestward dip on the Toro-Bunyoro Fault. Fault motions are predominantly normal, with strike slip motions in between the Ruimi-Wasa and Toro-Bunyoro Faults.

The observed depth distribution of seismicity peaks at 16 km and falls off sharply by 22 km. Strength envelope models show that heat flow in the range of 54 mWm\(^{-2}\) to 66 mWm\(^{-2}\) is consistent with the depth distribution of seismicity. This heat flow range is similar to heat flow in other mobile belts in East Africa away from major rift valleys and is lower than the high heat flow (~70-100 mWm\(^{-2}\)) reported for the Eastern Branch in Kenya and the Western Branch in Tanzania. A heat flow of 54 to 66 mWm\(^{-2}\) is also consistent with average heat flow from Precambrian terrains globally. Therefore, heat flow in the Western Branch around the northern nose of the Rwenzori Mountains is not elevated.

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References


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