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Near surface temperature lapse rate for treeline environment in western Himalaya and possible impacts on ecotone vegetation

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Abstract: This study presents the maiden results of near surface temperature lapse rate (TLR) for treeline environment in the western Himalaya in India based on ground observations along an elevation transect (1500–3680 m). Statistically significant correlation and linear regression model was used to calculate TLRs for different months. The mean annual TLR in western Himalaya is less steep (−0.53 °C/100 m) than the commonly used value (−0.65 °C/100 m). Notably, the lapse rates of temperature varied across different seasons and the two study aspects suggesting that TLR is governed by micro-climatic and physiographic features. The highest mean TLR (−0.64 °C/100 m on NW aspect and −0.60 °C/100 m on SE aspect) was observed for pre-monsoon season (March–May) whereas the lowest (0.42 °C/100 m on NW aspect and −0.39 °C/100 m on SE aspect) for the winter season (December–February). The annual cycle of TLR reveals a bi-modal pattern with two maxima in the pre-monsoon and post-monsoon seasons whereas two minima in winter and monsoon, respectively. The higher TLR in dry or warmer and lower in humid or cold atmospheric conditions suggest different controlling factors determine TLR in the individual seasons. There is a need to examine whether the low TLR of the present study transect (−0.53 °C/100 m) is because of elevation-dependent warming (being more in higher elevations) under the influence of global climate change. The observed shallower TLR, an indication of elevation-dependent warming, may have several implications on ecotone vegetation in Himalaya under changing climate scenarios.

Key words: Climate change, ecotone vegetation, elevation dependent warming, temperature lapse rate, treeline, Western Himalaya.

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Introduction

Alpine treeline ecotone, occurring between a subalpine forest and alpine meadow, is an extremely temperature-sensitive transition zone (Körner 1998). Sensitivity of plant species to change in temperature and other abiotic factors (e.g. radiation, moisture, wind, slope exposure, topography) is high across this ecotone (Holtmeier & Broll 2005; Körner & Paulsen 2004; Li *et al.* 2008; Wang *et al.* 2005). The alpine regions in Himalaya are considered climate hotspots and indicator zones of species geographic range shift induced by climate change and global warming (Grabherr *et al.* 1994; Körner 2003; Lesica & Steele 1996; Telwala *et al.* 2013; Walther *et al.* 2002). Over past few decades accelerated rates of warming are noticeable in most of the Himalayan regions (Bhutiyani *et al.* 2010; Gao *et al.* 2004; IPCC 2007; Joshi & Kumar 2013; Shrestha *et al.* 1999). A positive feedback effect, arising from decreasing albedo on account of decline in snow cover and change in land use-land cover, is one of the possible reasons for warming in high altitude mountainous regions (Pepin & Losleben

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2002; Rangwala & Miller 2012). The spatial patterns of warming have an unequivocal impact on ecological balance of mountain ecosystems which is largely influenced by local scale variations of climate parameters (Gerlitz *et al.* 2015). We hypothesize that elevation-dependent warming (Pepin *et al.* 2015) in mountain environments may alter the temperature lapse rate (TLR, defined as the rate of change in temperature with rise in elevation) of near surface temperature. The near surface TLR is a controlling factor of many environmental processes and hence the altitudinal dependence of temperature in mountainous region is important to study impacts of climate change on various processes (e.g. climate-vegetation interactions, glacier retreat and melt runoff, change in hydrological and moisture regimes, etc.).

Different values of environmental lapse rate (e.g. −0.55 °C/100 m, −0.60 °C/100 m, −0.65 °C/100 m) are often used to estimate air temperature at ungauged sites when low precision suffices (Rolland 2003). However, in complex Himalayan terrain, climate at regional and local scale is affected by topography, latitude, movement of air and vegetation patterns (Barry 1992). Because of these factors climatic parameters (e.g. temperature and precipitation) in this region vary remarkably even over a short geographic distance and hence the regional distribution of temperature and precipitation vary along altitudinal gradient. Lapse rates may vary with latitude, topographic slope, regions, and season (Bolstad *et al.* 1998; Dodson & Marks 1997; Immerzeel *et al.* 2014; Rolland 2003), hence cannot be treated as spatially and temporally constant. Therefore, simply adopting the general temperature-elevation (T - E) trends and the rough approximations or estimate may not effectively explain T - E trends at local level. Some studies across the globe have further confirmed that the temperature lapse rates and precipitation gradient are highly variable in space and time in mountain regions and its magnitude may vary in different location as a function of energy balance (Gardner *et al.* 2009; Gouvas *et al.* 2011; Kattel *et al.* 2012; Rolland 2003; Thayyen & Dimri 2014). Therefore, assumptions of persistent values of TLR in mountains is imprecise and may lead to erroneous results about change in elevation dependent warming rate and its impacts on different ecosystems along elevation gradient.

Several studies have been carried out in different mountain regions across the world (e.g. Washington Cascades Mountains, Northern Italy and Autrian Alps, Mt. Taibai, Colardo Rocky Mountains, and in the Northern Hemisphere) on variability of TLR and its controlling factors (Diaz & Bradley 1997; Minder *et al.* 2010; Pepin *et al.* 2015; Rolland 2003; Tang & Fang 2006). However studies on temperature gradient based on observed data are lacking for Himalayan regions which encompass very wide bioclimatic elevational gradients. This is mainly because of the fact that the ground based meteorological observations in Himalaya are very scanty and location specific long term precise climate records are limited, particularly for tree line environments and other high altitude ecosystems in Indian part of Himalaya (Friedland *et al.* 2003; Shrestha *et al.* 1999). Given this background, the key questions addressed in this paper are: (i) Does the near surface temperature lapse rate in tree line environment vary across the seasons?, (ii) Does the aspect play a role in distribution of the temperature and its elevation gradient in Himalaya?, and (iii) How does TLR for treeline environment in Himalaya differ from the global average? Apart from these, based on published literature, this study also highlights the possible impacts of observed TLRs on the treeline vegetation. As far as our knowledge goes, this is the first attempt to estimate temperature lapse rates for treeline environment in western Himalaya. In this study we hypothesize that (a) TLR should decrease because of elevation-dependent warming, which stresses more temperature rise in higher elevations than in lower elevation, (b) because of reduced snow cover and concomitance decrease in albedo, the elevation dependent warming could be higher in winters, (c) this difference could be reflected in lower TLR during winters, (d) in other words, climate change may alter TLR.

Study Site

The present study was carried out in and around treeline ecotone of Chopta-Tungnath transect in Indian Western Himalaya (30°30.76′–30°27.59′N; 79°05.54′–79°16.55′E) varies between 1600 and 3680 m and 1500 and 3680 m along South Eastern (SE) and North Western (NW) slopes, respectively (Fig.1a). The forests at the higher elevations of the study area fall under the subalpine zone, which gives way to alpine meadows beyond the timberline ecotone (Rai *et al.* 2012). In higher altitudes, above treeline (3400–3700 m) region of the study transect, the grasslands are dominated by herb species of *Anemone, Potentilla, Aster, Geranium, Meconopsis, Primula* and dotted pockets of shrubs of *Rhododendron anthopogon* and *Juniperus* species. Whereas, in sub-alpine region (i.e. between 2900–

(a) (b)

Fig. 1. (a) Location map of the installed loggers within the study area (Source: Google Earth), (b) Automatic Weather Station installed at a high altitude site (3360 m; $30^0$29.57' N, $79^0$12.95'E).

3400 m) mixed forests dominated by species like *Rhododendron arboreum, R. campanulatum* scattered with a few *Abies pindrow* and *Taxus baccata* trees are present. Below 2800 m, broad-leaved forests are dominated by *Quercus semecarpifolia*, *Betula utilis, Abies spectabilis, Acer caesium, Rhododendron arboreum* and *Sorbus foliolosa.* The climate of Tungnath region is characterized bylong wet summers and long and cold winters. Based on observed meteorological records obtained from an AWS installed within the study area at 3360 m asl (30°29.57′N, 79°12.95′E; Fig. 1a,b), the average precipitation and temperature for 2017 were found as 2941.59 mm and 6.03 °C respectively. The climate of the Chopta-Tungnath transect is dominated by monsoon circulation, with pre-dominant easterly winds in the summer and westerly winds from October to March.

Methods

Meteorological Setup and Data

To estimate the TLR and its spatial variations over the study area, 10 portable ONSET HOBO Pro-V2 microbloggers and 6 RG-200 tipping bucket rain gauges (Global Water make; 8" dia.) were installed along an elevation gradient from 1500 m to 3680 m at two different aspects (N-W & S-E). The temperature loggers were covered with radiation shields to protect the sensors from direct incoming shortwave radiation on exposed sunlight. On the N-W slope, a more advanced AWS was installed at an altitude of 3360 m asl which consists of a Campbell SR50A sonic ranging sensor, an ARG-100 tipping bucket with a simple HOBO event data logger, and a Campbell temperature probe (109-L), and other sensors for measurement of temperature, soil moisture, wind speed and direction, and net radiation. Station data from all the sensors installed at different locations were recorded at 15 minutes intervals and the data collected during one-year period (December 2016 to November 2017) were used for various analyses.

The data were analysed for the entire year (December 2016 to November 2017) and seperately for the four main climatic seasons, namely Winter (December-February), Pre-monsoon (March–May), Monsoon (June–September), Post-monsoon (October–November). These distinct seasons were identified based on analysis of the observed precipitation and temperature data from two high altitude station located at 3360 m asl and 3140 m asl on two different aspect within the study transect and also based on the literature (Immerzeal *et al.* 2014; Kattel *et al.* 2012) for comparison of results. Fig. 2 show variation in daily temperatre and precipitation at a high altitude station (3360 m asl) on NW slope where approximately 92% of the total auunal rainfall occured during monsoon and nearly 6% occured during post-mosoon season. Similar patterns of variation in temeprature and precipitation are

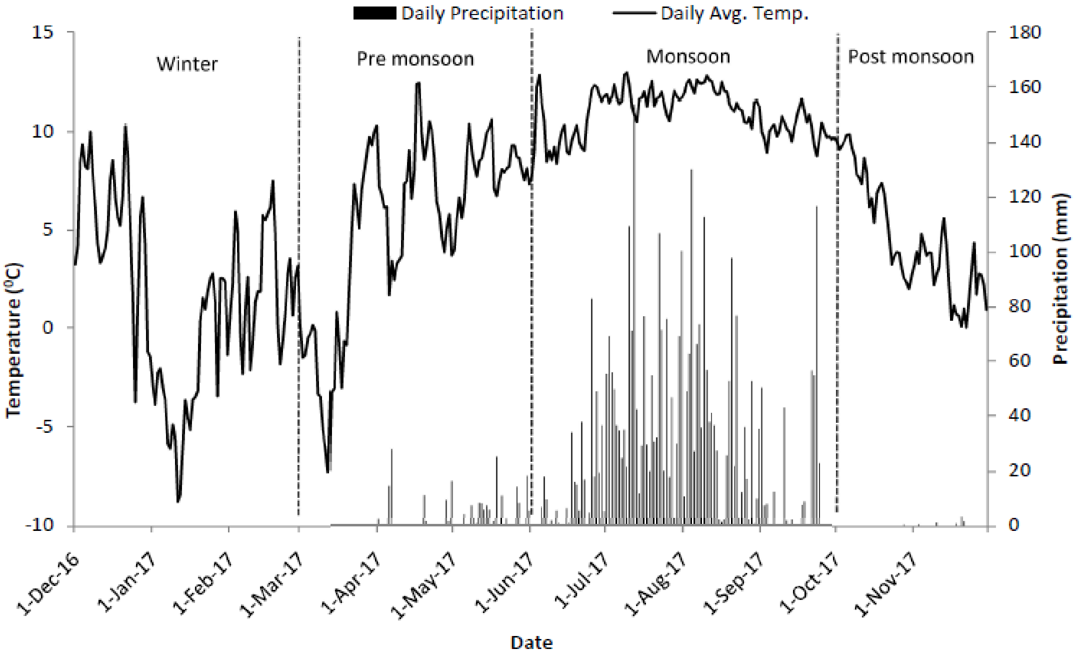

Fig. 2. Daily rainfall histogram and mean temperature at high altitude station (3360 m asl) on NW-aspect.

observed for SE slope; approximately 86% of the total aunnal rainfall occured during monsoon whereas nearly 12% occured during post-mosoon season. Within the study transect, the winter season at high altitudes is chareterized by the lowest temperatures and precipitation mostly in the form of snow. The pre-monsoon season is chareterized by relatively high temperatures gradually increasing till the onset of monsoon. During this period, high variability in diurnal temperature and small events of precipitation (average rainfall 315 mm) are obsrved. During the monsoon season heighest amount of rainfall (2120 mm) is received along with relatively high temperature, and charecteristically low diurnal varibility. A steady decrease in temperature alongwith a negligible amount of rainfall (average 7 mm) is the key feature of the post monsoon (autumn) season. Mean temeparture and precipitation records from two study aspects, north-west (WE) and south-east (SE) for the four seasons are summarized in tables 1 & 2.

Data analysis

The correlation coefficients between temperature and elevation of the station was calculated for the observed dataset. Monthly mean, maximum, and minimum temperatures were correlated to elevations, and the significant correlation coefficients were presented (Table 3). Correlation analysis was carried out to estimate the measure of the strength of the linear relationship between the two variables and test how strongly temperature is linearly controlled by elevation. The Pearson Correlation's Coefficient (r) between the two variables Temperature (T) and Elevation (z) can be expressed in the following forms:

$$r(T,Z) = \frac{1}{n}\sum_{i=1}^{n}\left(\frac{T_i - \bar{T}}{S_T}\right)\left(\frac{z_i - \bar{z}}{S_z}\right)$$

Where $\bar{T}$ and S_T, respectively, represents mean and standard deviation for temperature whereas $\bar{Z}$ and Z_T are the mean and standard deviation values for elevation. If the relationship between the variables is not linear, then the correlation coefficient does not adequately represent the strength of the relationship between the variable.

All temperature data were then aggregated to hourly values for various analyses. Generally, air temperature is assumed to decrease/increase linearly with elevation under well-mixed atmospheric boundary conditions (Lundquist *et al.* 2008). Temperature lapse rates were calculated by developing a regression equation using all point level observations of temperature and elevation. The developed regression determines nature of linear association between the two variables. TLRs (°C/100 m) were estimated using following regression equation:

$$LR = \frac{T_1 - T_2}{z_1 - z_2} = \frac{dT}{dz}$$

Table 1. An overview of the mean temperature (T in ^{0}C) and total rainfall (P in mm) for four seasons on North-West (NW) aspect.

Season	Station name (altitude in m asl)								
	Ukhimath (1500)**		Tala (1820)*	Dugalbhitha (2500)**		Chopta (2870)*	HAPPRC (3360)***		Chandrashila (3680)*
	T	P	T	T	P	T	T	P	T
Winter	12.61	---	9.69	5.57	---	3.27	1.72	---	2.48
Pre-monsoon	18.33	287.93	15.15	11.29	358.5	9.26	5.43	219.96	4.30
Monsoon	22.49	1425.49	19.48	16.07	2067.61	14.91	10.93	2721.62	10.19
Post-monsoon	17.07	0.5	14.30	9.88	5.31	8.19	4.56	10.41	6.57
Annual	17.6	1714.0	14.6	10.7	2431.4	8.9	5.6	2953.0	5.8

*Temperature logger, **Temperature logger & Raingauge, ***Automatic Weather Station.

Table 2. An overview of the mean temperature (T in ^{0}C) and total rainfall (P in mm) for four seasons on South-East (SE) aspect.

Season	Station name with altitude (in m asl)							
	Sirauli (1600)*	Mandal (2100)**		Kanchula Khark (2675)**		Saukhark (3100)**		Chandrashila (3680)*
	T	T	P	T	P	T	P	T
Winter	11.6	10.5	---	4.2	---	5.2	---	2.9
Pre-monsoon	18.1	15.6	280.47	9.8	379.8	7.7	371.1	4.3
Monsoon	22.4	20.1	1920.68	14.8	2208.8	12.7	2394.3	10.2
Post-monsoon	16.5	14.2	7.75	8.3	9.6	7.2	4.5	6.6
Annual	17.2	15.1	2209.0	9.3	2598.3	8.2	2770.0	6.0

*Temperature logger, **Temperature logger & Raingauge, ***Automatic Weather Station.

Table 3. Correlation between maximum, minimum, mean temperature and elevation

Months	Maximum temperature		Minimum temperature		Mean temperature	
	TLR (°C/100 m)	*r*	TLR (°C/100 m)	*r*	TLR (°C/100 m)	*R*
January	–0.67	–0.94*	–0.45	–0.94*	–0.53	–0.95*
February	–0.59	–0.93*	–0.42	–0.96*	–0.52	–0.97*
March	–0.82	–0.99*	–0.44	–0.95*	–0.61	–0.98*
April	–0.82	–0.97*	–0.51	–0.98*	–0.66	–0.99*
May	–0.84	–0.96*	–0.54	–0.99*	–0.67	–0.99*
June	–0.79	–0.97*	–0.51	–0.99*	–0.66	–0.99*
July	–0.62	–0.98*	–0.47	–0.99*	–0.54	–0.99*
August	–0.60	–0.98*	–0.44	–0.98*	–0.52	–0.98*
September	–0.63	–0.97*	–0.44	–0.98*	–0.55	–0.98*
October	–0.51	–0.95*	–0.47	–0.97*	–0.52	–0.97*
November	–0.30	–0.89*	–0.38	–0.94*	–0.36	–0.94*
December	–0.34	–0.75**	–0.18	–0.83**	–0.20	–0.83**

*Correlation values significant at $P < 0.01$; *correlation values significant at $P < 0.05$.

Where T_1 and T_2 are the air temperature of the highest and lowest points (in °C), and z_1 and z_2 are their respective elevations (m). The TLRs were calculated from regression of all values; this allows calculating the strength of the relationship between air temperature and elevation.

Considering the fact that temperature in mountains is regulated by topography, aspect and

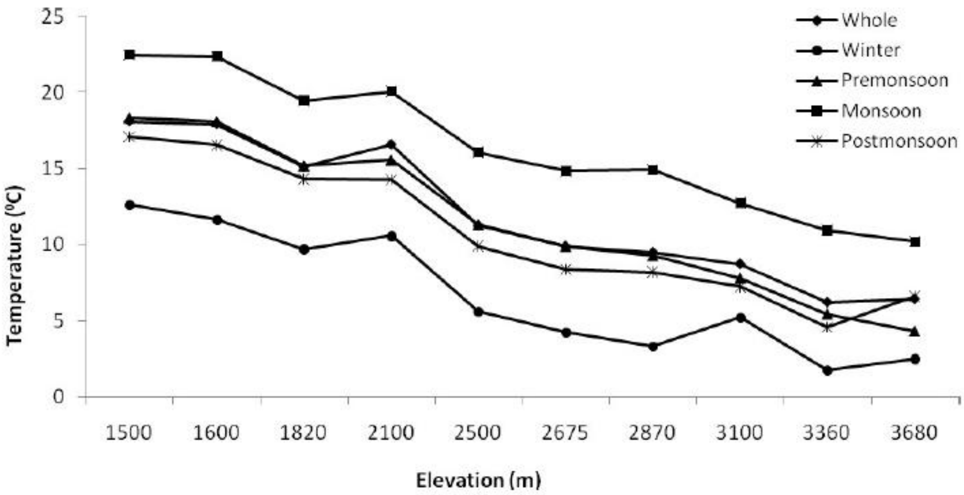

Fig. 3. Mean temperature versus elevation plot for entire transect for all four season and the whole year.

seasonal patterns of climate (Barry 1992), TLRs were calculated for the entire year as well as separately for each of four seasons; namely, pre-monsson (MAM)), monsoon (JJAS), post-monsoon (ON), and winter (DJF) for the study aspects seperately. Statistical analyses were performed to test the significance of results obtained. First, the significance of difference between the TLR for mean, maximum and minimum temperatures on two different aspects was tested using the independent samples t-test. An independent-sample *t*– test was performed to determine if the estimated TLR of each aspect was different from a general lapse rate value. Second, one-way analysis of variance (ANOVA) was used to compare the seasonal TLRs of one aspect to that of the other for the study transect, iteratively. The t-test and ANOVA were evaluated based on the significance alpha level (α = 5% or 95% confidence interval) and p values ($P < 0.05$) using STATISTICA 8.0 statistical package.

Results

Correlation analysis between elevation and temperature

The correlation between monthly temperature and elevation was significantly negative at $P < 0.01$ and $P < 0.05$ for all months (Table 3). Among the seasons the lowest correlations(r varies between 0.75 to 0.95) were during winter months when temperatures were the lowest. Value of correlation coefficient increased from March and remained at the heighest level from May to September. Relatively lower correlation coefficients exists between maximum temperature during summer and elevation on account of the frequent terrain-dependent monsoon rainfall events (Kattel *et al.* 2012).

Temperature lapse rate variation in treeline line environment

Increase in temperature from winter to monsoon is obvious across the entire transect (Fig. 3). Significant correlations (at $P < 0.01$ and $P < 0.05$) indicate that the relationship between temperature and elevation is strong (Table 3); hence tempe-rature can be accurately predicted as a function of elevation. The TLRs for different months calculated from corresponding linear regression are presented graphically in Figs. 4 & 5.

Monthly variation in temperature lapse rate

Monthly variation in lapse rates for the two aspects (i. e. NW & SE) is shown in Figs. 4 & 5. Higher negative TLR values than seasonal mean are considered as the greater decline in temperature with increasing elevation whereas the less negative values of TLR are considered as the temperature inversion (Kattel *et al.* 2012). Along NW aspect, the mean TLR varied from −0.23 (± 0.55) °C/100 m in December to −0.69 (± 0.20) °C/100 m in May (Fig. 4). For maximum temperature, the highest lapse rate (−0.89 ± 0.77 °C/100 m) was observed in May and the lowest (−0.34 ± 0.01 °C/100 m) in November. TLR for minimum temperatures ranged from −0.18 (± 0.65) °C/100 m in January to −0.55 °C/100 m in May and June.

Along SE slope, the lowest TLR for mean temperature is observed in December (−0.18 ± 0.37

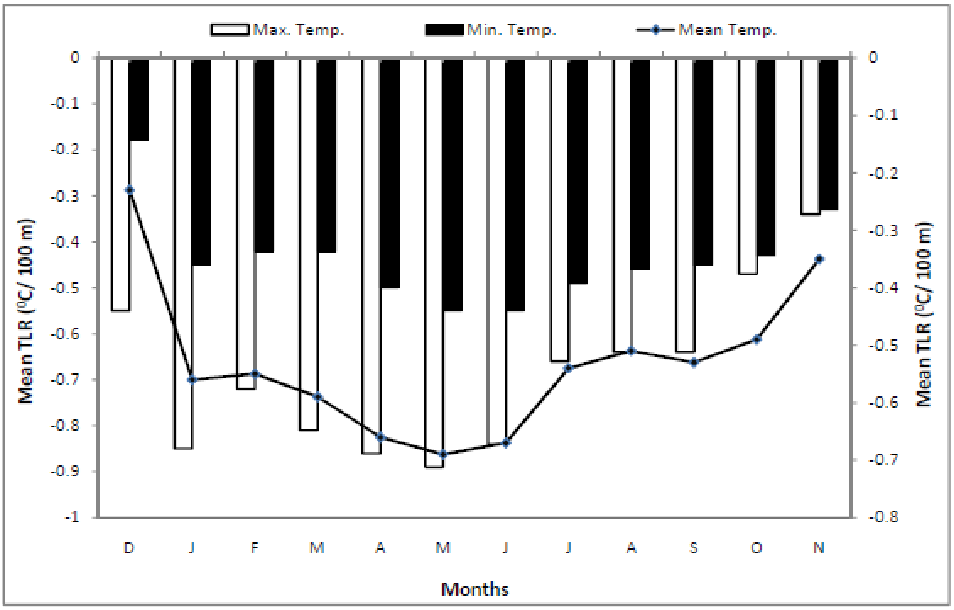

Fig 4. Variation in monthly temperature lapse rates along NW aspect of the study transect.

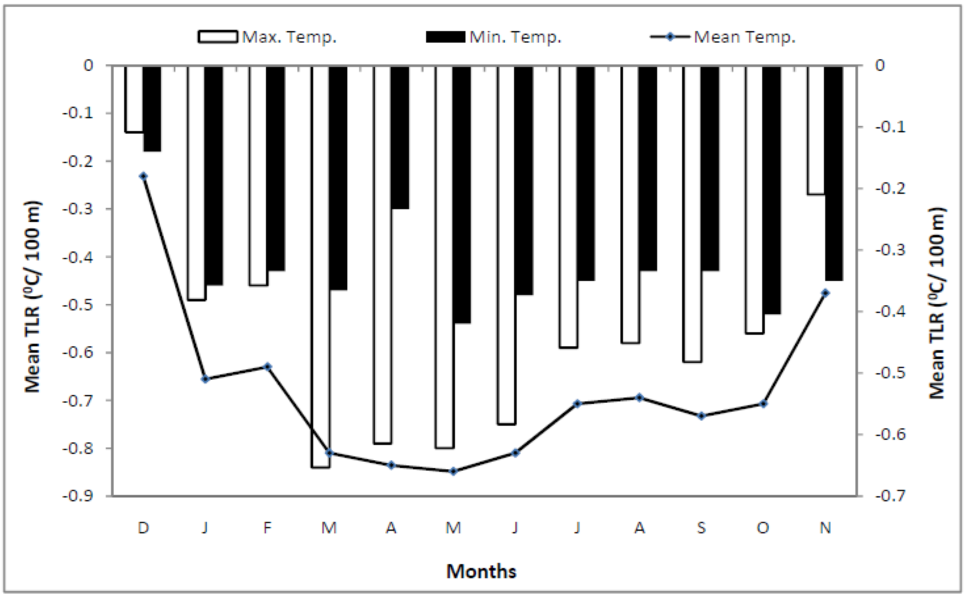

Fig. 5. Variation in monthly temperature lapse rates along SE aspect of the study transect.

°C/100 m) and highest in May (−0.66 ± 0.20 °C/100 m). For maximum temperatures, the TLRs were found to be varying between −0.14 ± 0.13 °C/100 m in December to −0.84 ± 0.42 °C/100 m in March. Whereas, for minimum temperatures, TLR varied between −0.18 ± 0.37 °C/100 m in December to −0.54 ± 0.02 °C/100 m in May (Fig. 5). Results show that TLR for maximum temperatures were significantly higher than that for minimum temperatures both for NW aspect ($t = 6.553$; $P = 0.00004$) and SE aspects ($t = 2.725$; $P = 0.019$).

Seasonal and Annual variation in TLRs and role of aspect

The highest TLR was observed for pre-monsoon season and the lowest during winter season for both mean and maximum temperatures. However, for

Table 4. TLR (in °C/100 m) for mean, minimum and maximum temperatures in relation to aspects and seasons.

Season	N-W Aspect			S-E Aspect		
	Mean TLR	Maximum TLR	Minimum TLR	Mean TLR	Maximum TLR	Minimum TLR
Winter (DJF)	–0.42	–0.71	–0.35	–0.39	–0.36	–0.35
Pre-monsoon (MAM)	–-0.64	–0.85	–0.49	–0.60	–0.81	–0.51
Monsoon (JJAS)	–-0.57	–0.67	–0.55	–0.55	–0.63	–0.57
Post-monsoon (ON)	–-0.44	–0.41	–0.38	–0.47	–0.42	–0.48

minimum temperatures, the highest TLR occurred during monsoon season on both the aspects (Table 4). Steepest average TLR is observed during pre-monsoon season (−0.64 °C/100 m on NW aspect and −0.60 °C/100 m on SE aspect) whereas lowest mean TLR was observed during winter season (−0.42 °C/100 m on NW aspect and −0.39 °C/100 m on SE aspect). Between the two aspects, TLR values were higher for NW aspect than for SE aspect. A significantly lower TLR was observed on SE aspect than on NW aspect for mean temperatures ($t = 2.06$; $P = 0.035$) and minimum temperatures ($t = 2.1483$; $P = 0.04$), but not in case of maximum temperatures ($t = 1.44$; $P = 0.16$). The two aspects also differed significantly ($P \leq 0.05$) in seasonal distribution of TLR (Table 4). On an annual scale, a higher value of mean TLR was observed for NW aspect (−0.46 °C/100 m) than the SE aspect (−0.36 °C/100 m). Based on the analysis carried out using One-Way ANOVA at 5% level of significance, it was found that the TLR varied significantly across all the seasons for the mean ($F = 3.2175$; $P = 0.03$), maximum ($F = 3.675$; $P = 0.019$), and minimum ($F = 9.895$; $P = 0.0002$) temperatures on both the aspects indicating that the seasons have strong influence in determining the TLR across the study transect and in general in the Himalaya. Our results show that high variability exists in temperature lapse rates for different months, seasons and aspects.

The analysis for annual cycle of TLR for maximum, minimum, and mean temperature for both the aspects shows that the seasonal variations of TLRs exhibit bi-modal patterns with peaks in pre-monsoon and post-monsoon periods and with two lowest values in the winter and monsoon seasons (Figs. 6 & 7).

Discussion

The mean TLR values (−0.53 °C/100 m) estimated in the present study based on observed data of an elevational transect is lower than the used in past literatures (−0.65 °C/100 m). It indicates that the temperature in higher ranges would be warmer than estimated on the basis of earlier used TLR.

Relatively lower winter time correlation coefficients may be related to temperature inversion effect that weakens the temperature-elevation relationship (Marshall *et al.* 2007). Opposite to the TLR, the inversion effect sets up when warmer air is displaced by the sinking cold air, particulary during night time (Rolland 2003). The higher LRs for maximum tempetaure than that for minimum temperature on both the aspepcts of the study transect may be due to the adiabatic mixing within the boundary layer during day time (Kattel *et al.* 2012).

The higher values of TLRs observed for NW aspect in comparison to SE aspect could be related to more rainfall on SE aspect than NW aspect; based on the data recorded from all the stations, the average rainfall for monsoon season was found as 2175 mm and 2070 mm on SE and NW aspects, respectively. Higher LR values in drier conditions than humid conditions have also been reported by Immerzeel *et al.* (2014), Kattel *et al.* (2015) and Tang & Fang (2006). The lapse rates observed in the present study support our hypothesis that the TLRs are controlled by both variation in topographic features (slope and aspect) and climatic conditions (relative humidity).

The highest mean TLR and the difference of lapse rate between maximum and minimum temperature were observed during pre-monsoon season on both the aspects (Table 4). Because of the absence of clouds and higher temperature during pre-monsoon, land surfaces receive more incoming solar radiation compared with outgoing radiation. This phenomenon results in a rise in daytime surface temperature and large sensible heat flux which can enhance strong dry convection in the day-time (Blandford *et al.* 2008). The pre-monsoon season also has the highest daytime saturation varour

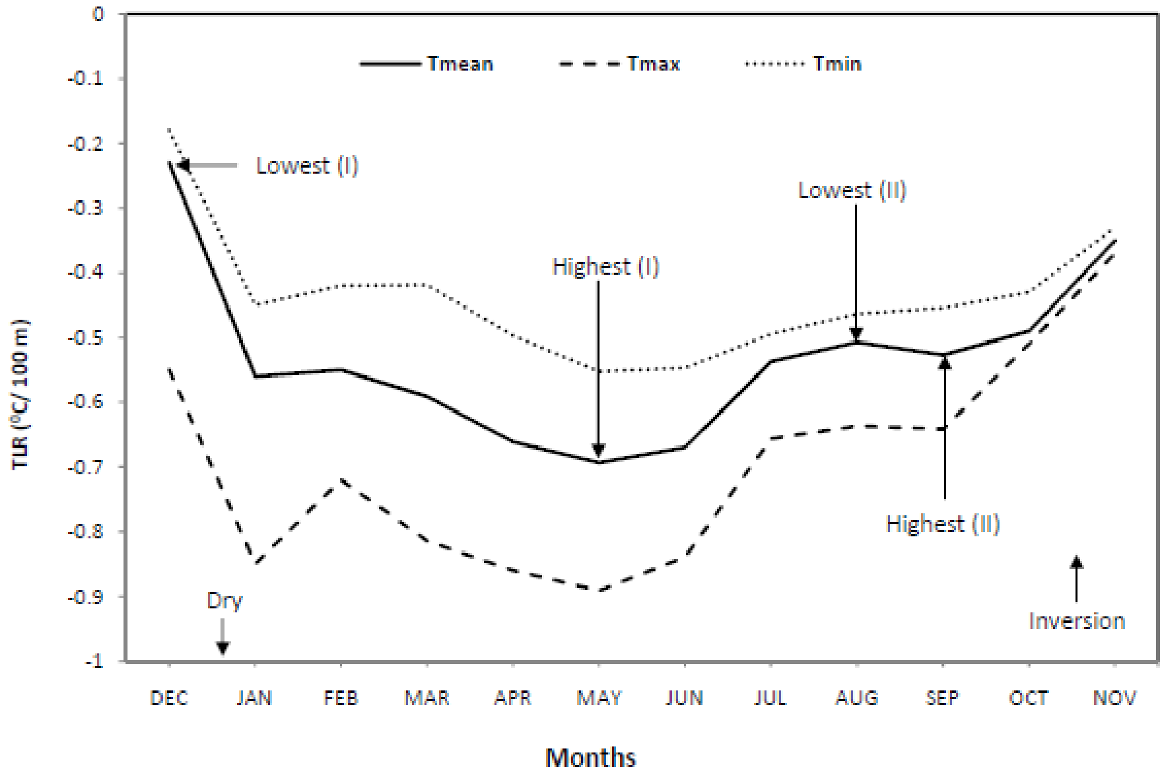

Fig. 6. Annual cycle of maximum, minimum, and mean TLR on NW aspect.

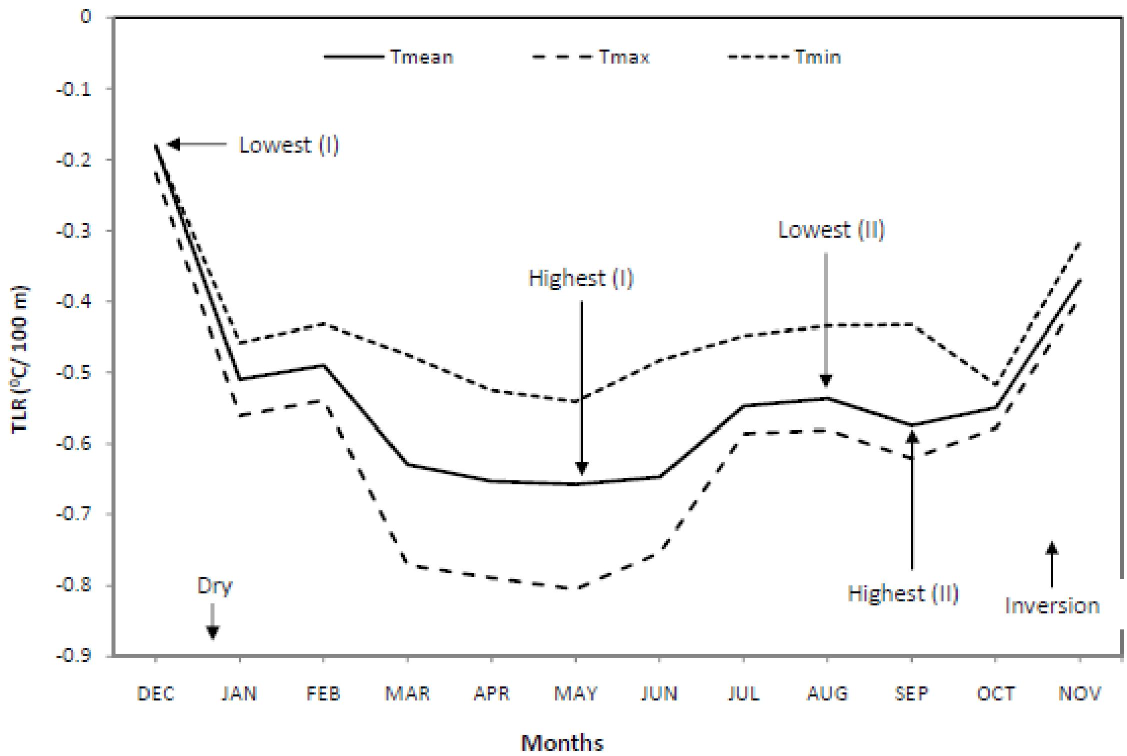

Fig. 7. Annual cycle of maximum, minimum, and mean TLR along SE aspect

pressure (e_s) lapse rate (Kattel *et al.* 2012); as a result the TLR reaches a maximum value in this season.

The second TLR peak was observed during the post-monsoon season (Fig. 6 & 7). The magnitude of mean TLR in this season was relatively lower compared with the pre-monsoon season but higher in comparison to the winter and pre-monsoon seasons. The post-monsoon season also had the least difference between the maximum (−0.41 °C/100 m for NW slope and −0.42 °C/100 m for SE slope) and minimum (−0.38 °C/100 m for NW slope and −0.48 °C/100 m for SE slope) TLR (Table 4). Weather conditions during this season are similar to pre-monsoon conditions except that the thermal forcing effect is relatively small due to the turbulence between the two climatic phases, i.e. retreat of monsoon and the onset of winter westerlies. Our results are in close proximity with the TLR values and variations in them with those of western and central Himalaya in Nepal (Immerzeel *et al.* 2014; Kattel *et al.* 2012, 2015).

The lowest mean TLR values were observed during winter season (DJF) on both the aspects. The lowest value of TLR in winter implies that other controlling factors play a more important role. During winter, sky mostly remains clear and radiative cooling is intense, leading to a stable stratification and a temperature inversion. This condition facilitates the development of suitable microclimates (Thyer 1985) for cold air deposition in low areas through down slope flow at the flat terrain and lowland valleys. The temperature inversion may be further enhanced by foggy conditions during the winter season (Rolland 2003). While reduced snow cover in the higher parts of transect is expected to reduce albedo and raise temperature, the increased aerosol concentrations in lower elevation is likely to cause cooling. These opposite trends are likely to reduce winter TLR.

The second lowest TLRs were observed during the monsoon season (JJAS) when moisture content in the air is high throughout the elevation range. This is because the air over high altitude areas is warmed by latent heat release associated with water vapor condensation, thereby rendering temperature uniformly high. Latent heat transfers depend on moisture content in the atmosphere (Marshall *et al.* 2007). Hence, in response to heavy summer rainfall, surface temperature decreases and moist adiabatic processes prevail (Thyer 1985). Thus, the stable lapse rate for both maximum (−0.67 °C/100 m) and minimum (−0.55 °C/100 m) temperature in monsoon implies a strong association with rainfall amount. This indicates that the annual cycle of varour pressure (e_s) plays a contributory role in enhancing and reducing the TLRs during dry (in pre-monsoon season) and moist conditions (in monsoon season), respectively. Furthermore, TLR variability is also governed by the cloudiness. The thick cloud cover during the monsoon months (JJAS) leads to reduction in insolation during the day (Kattel *et al.* 2012) and detention of outgoing long-wave radiation during the night time (Bhutiyani *et al.* 2007), leading to higher minimum surface temperatures regardless of elevation within a limit.

The bi-modal pattern of TLR variability observed in treeline environment of western Himalaya is distinct from other mountain regions such as south-central Idaho, Washington Cascades, Colorado Rocky Mountains, and semiarid southeastern Arizona in the USA; Northern Italy, Greece, and the Austrian Alps in Europe; and northern and southern slope of Mt. Taibai in China (Blandford *et al.* 2008; Diaz & Bradley 1997; Gardner *et al.* 2009; Harlow *et al.* 2004; Marshall *et al.* 2007; Minder *et al.* 2010; Pepin & Losleben 2002; Richardson *et al.* 2004; Rolland 2003; Tang & Fang 2006). Studies carried out in these regions have reported lower TLR in winter and higher TLR in summer due to the maximum dry convection in summer and minimum in winter on account of temperature inversion. Majority of these study regions are located in mid-latitude mountain systems, where there is no distinct summer wet season or where winter is the wettest season. The precipitation pattern of Himalayan region is determined by two atmospheric circulation patterns: one driven by the Indian summer monsoon or south-easterlies and the other driven by the westerlies in winter (Singh *et al.* 2017). Contribution of these two characteristics in Himalayan region might contribute to a distinct bi-modal pattern of TLR over the study transect. The lower values of LR of our study in comparison to those of the past might be due to enhanced elevation dependent (EDW) warming in Himalaya. Throughout the world warming is being observed more in higher elevation than lower elevation **(**Pepin *et al.* 2015). Such a differential warming is expected to decrease the TLR as a consequence of EDW under the influence of climate change.

Implications of low TLR for timberline elevation

Low TLRs particularly during winter and monsoon months are likely to have contributed to high treeline in Himalayas. Heat deficiency is considered to be the main cause of treeline formation (e.g., Körner 2012). Winters in Himalayas are mild (Sakai & Malla 1981) partly because day lengths remain long (> 10 hr), and days are sunny (Zobel & Singh 1997). As for growth of trees in high elevations, monsoon (warm and moist) is the key period, when much of the net primary productivity occurs (Singh & Singh 1992). Hence elevation dependent warming is likely to be now a major contribution to elevational rise of treeline.

Conclusion

This study presents initial estimate of TLR for treeline environment in western Himalaya. Our estimate of TLR based on observed data (−0.53 °C/100 m) is distinctly lower than values of TLR used in the past for Himalayan region. Since EDW results in a decrease in temperatures between low and high elevations, and is likely to be connected with global climate change, the low TLR of our study (−0.53 °C/100 m) and that reported recently in Nepal for Langtang valley (Immerzeel *et al.* 2014) may be a consequence of global warming. Our study establishes that TLR varies seasonally; it is generally lower during winters and monsoon period whereas higher during pre-monsoon and post-monsoon periods. This study has shown for the first time that TLR varies from one aspect to another, which may partially explain the aspect-related difference in treeline elevation in Himalayas (Schickhoff 2005).

The low TLR indicates that conditions in high Himalayas are likely to be warmer than generally held out. The warmer temperatures as such would promote plant growth, provided the increased evapotranspiration loss does not become limiting. The lower values of TLR may have several possible impacts on the dynamics of treeline ecotone in Himalaya. The change in snow and moisture regime, increased evapotranspiration and water stress, change in albedo and surface energy balance, and modified distribution patterns, and shift in range and growing season of alpine vegetation are some example. Considering observed variability in TLR, the present study advocate for use of seasonally varying lapse rates to assess impact of climate change on different ecosystems along altitudinal gradient. Further, information about changing climate and vegetation patterns of climatically sensitive alpine ecosystems is very crucial for a comprehensive understanding of their current and future states. The low TLR may partly explain the high treeline in Himalayas.

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