Effect of altitude and surface albedo variability on global UV-B and total radiation under clear-sky condition

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Abstract: Effects of snow cover and altitude on clear-sky global ultraviolet (UV-B) and total (G) solar radiation were studied at two positions in the High Tatras (Slovakia) in the period 2002–2004. The investigated sites are located at altitudes 810 m a.s.l. (Stará Lesná - SL) and 1778 m a.s.l. (Skalnaté Pleso – SP).
The UV-B irradiance corrected for equal total ozone content 300 DU was used to compare UV-B data observed during a period with snow cover and without any snow. The radiative transfer model TUV (Total ultraviolet-visible model) was used for UV-B data correction to uniform total ozone value. A relative increase in the UV-B irradiance caused by snow cover of 12-13% was found at SP, and of 13-16% at SL for SZA in the range of 55° - 65°. The total irradiance increase induced by snow of 12-18% and of 10-14% was determined at SP and SL respectively (SZA in range 55° - 65°).
To isolate the effect of remote snow area on investigated irradiances, the condition with continuous snow cover at SP and that of no snow at foothill position were studied separately. An increase in global irradiance of 7-10% was determined at SL in comparison with situations with snow-free surface at both observatories. No significant increase in the UV-B irradiance was found in this case at SL.
The dependence of studied irradiances on altitude was determined separately with respect to snow presence at compared places. For no-snow condition, the UV-B irradiance increase of 15% /1000 m was found, while the altitudal gradient of total irradiance was only 6% /1000 m. The UV-B fraction in the total solar radiation increases with altitude – the ratio UV-B/G is of 9% higher at SP by no-snow condition. During the spring months, when continuous snow persists at high altitudes, vertical gradient of the UV-B irradiances increased to 23% /1000 m.

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1. Introduction

A prolongation of snow cover in spring to summer periods, when the Sun culminates at small zenith angles, sharp differences between altitude of horizontally close places, complexity of terrain and typical regime of clouds are phenomena significantly affecting variations of solar radiation in mountain areas. A UV, especially the UV-B, radiation has strong biological action on live organism. It is important to determine the solar UV radiation variations in mountain areas, not only due to increased leisure activities at high altitudes (skiing, hiking), but also due to possible impact of mountains (especially due to longer lasting of snow cover) on the UV irradiances observed at lower, more densely populated valleys.

Although the effect of total ozone on the solar UV radiation variations is well established, there are still differences between vertical gradients and the UV effective albedos determined at different places.

The increase in solar radiation with altitude is caused by air molecule and aerosol particle number decrease with height. The attenuation of solar radiation by both noted phenomena is wavelength and SZA dependant (*Gröbner et al., 2000*). The vertical gradient of the UV radiation also depends on the air turbidity and on the UV-absorbing substances content in the layer between investigated places.

The relation between the solar radiation and the surface reflectivity is also wavelength and SZA dependant. The increase in effective albedo of the UV radiation is observed for a large SZA. *Degünther et al. (1998)* proposed that it is necessary to pay attention to the reflectivity variances in an area with radius 40 km, as it is an important factor determining the solar UV fluxes. Other authors proposed significantly lower radius of importance for the UV reflectivity.

Different values of the effective albedo in UV range have been determined at different places. The snow cover, its freshness and thickness have significant influence on the surface albedo variations.

Additionally, the complexity of terrain and the orientation of investigated areas affect the solar irradiation in mountains, too. In some cases, multiple reflections of radiative fluxes can enhance the observed UV-B irradiance (*Renaud et al., 2000*), but the effect of shading can diminish irradiances at mountain area in comparison with flat terrain (*Thomas, 1997*).

The aim of this paper is to investigate the effect of altitude and snow cover on the UV-B irradiances in comparison with the effect of noted phenomena on the total solar radiation in the High Tatras (Slovakia).

2. Material and methods

The effects of snow cover and altitude on the clear-sky UV-B, total (G), diffuse (D) and reflected (R) solar radiation was studied at two positions in the High Tatras (Slovakia) in the period 2002-2004. Both the investigated places are located at a south oriented slope of the Lomnický massive at altitudes 810 m a.s.l. (Stará Lesná - SL, 49.15 N, 20.28 E) and 1778 m a.s.l. (Skalnaté Pleso - SP, 49.18 N, 20.18 E). The vertical distance between these places is nearly 1 km (968 m), horizontal distance is about 4 km.

The UV-B irradiance was measured by temperature stabilized broadband UV-B radiometers (UV- biometers). A relative spectral response of the UV-B radiometer is close to biological action spectrum declared for human erythema (*McKinlay and Diffey, 1987*). The maximal relative spectral response is at wavelengths close to 290 nm. The relative spectral response determined by manufacturer slightly differs between SP and SL UV-B radiometers. The relative angular response of the devices is not known. This causes, that for large SZA (SZA $> 70°$), irradiances measured by both devices significantly differ due to the different instrumental characteristics. Both the UV-B radiometers are regularly compared with the Slovak standard instrument. A calibration constant of the devices is not corrected for equal total ozone and aerosol content. Changes of instrumental constants within $\pm 2\%$ were found for both UV-B radiometers during investigated period 2002-2004.

The total and scattered solar radiation was measured by the pyranometer Sonntag. A shadow-band was applied on the pyranometer to register the diffuse irradiance. The albedometer Janishevski was used for measuring

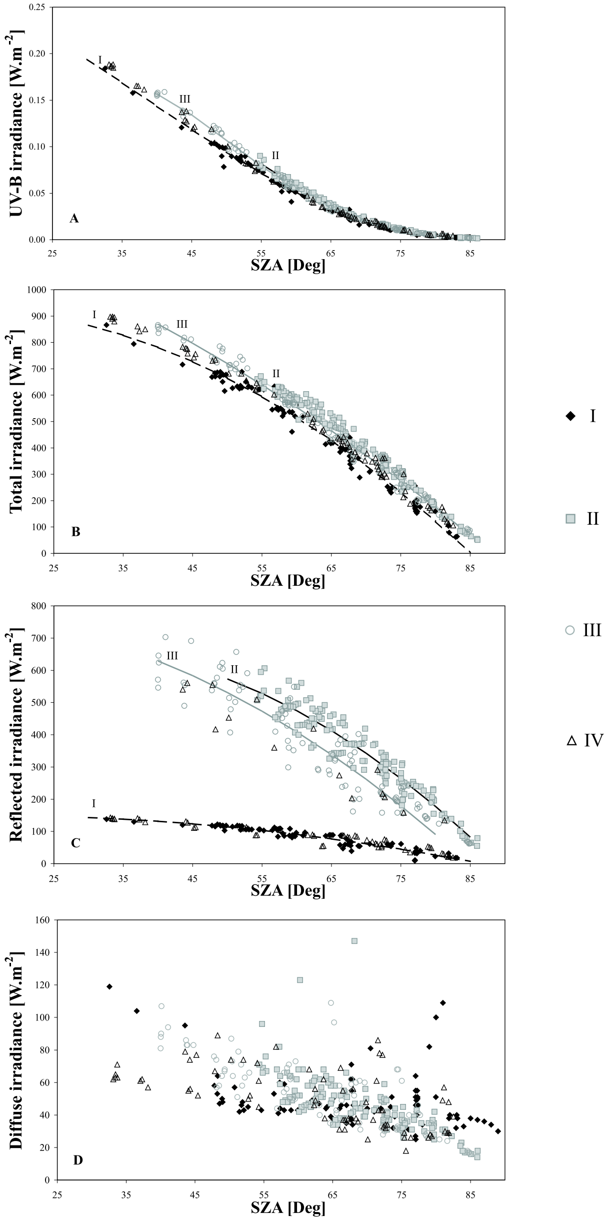

Fig. 1. The UV-B irradiance corrected to total ozone 300 DU (A), total (B), reflected (C) and diffuse (D) irradiance as a function of SZA for data groups I–IV at SP in the period 2002-2004. A dependence of irradiance on the SZA is expressed by polynomial function for data groups I-III (see text) and the UV-B, total and reflected irradiances.

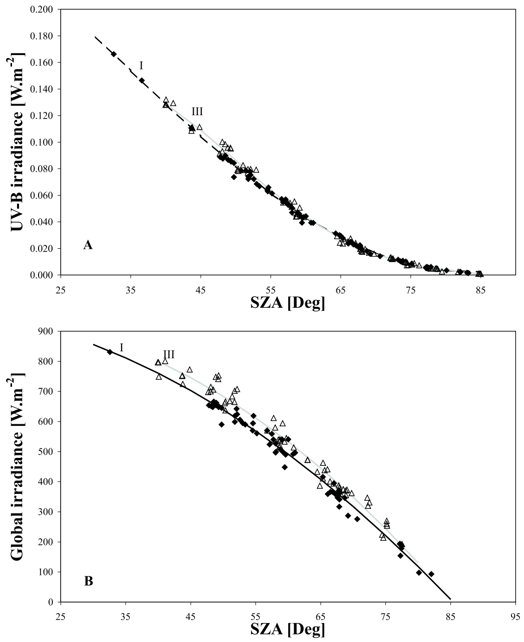

Fig. 2. The UV-B irradiance corrected to total ozone 300 DU (A) and total irradiance (B) as a function of the SZA for data groups I (diamonds) and III (triangles) at SL in the period 2002-2004. The irradiance data were fitted by polynomial function.

reflected solar radiation. Calibration of pyranometers is performed using the Link-Feussner actinometer and the Sun-shade method. The sunshine duration was registered by the Campbell-Stokes heliograph with blue tape.

Hourly averages of all registered parameters were elaborated. The SZA corresponding to hourly radiation data was calculated to time, respective to the center of every hourly interval. Additional information on weather, cloudiness and surface was also recorded every hour.

Comparing the summer and winter measurements of the solar radiation, it is necessary to take into account the effect of different Sun-Earth distances. The solar radiation data were corrected for the mean Sun-Earth distance (*Paltridge and Platt, 1976*). The difference between corrected and uncorrected UV-B no-snow data polynomial fit was 0–1% for $55° <$ SZA $< 65°$ at both stations. The corrected and uncorrected snow data fits distinguished less than 0 – (–1%) and (–1) – (–2)% for SL and SP, respectively.

The difference between relative change in the UV-B irradiance produced by snow cover calculated for the data corrected for the mean Sun-Earth distance and uncorrected data was 0 – 3% for $55° < SZA < 65°$. The difference was negligible in comparison with the polynomial fit uncertainty, but an omission of this correction can lead to the bias in snow effect determination, since snow data are obtained by lower Sun-Earth distance and snow-less data.

The UV-B irradiance corrected for the equal total ozone content 300 DU was used to compare the UV-B data observed during the period with snow cover and without any snow. The radiative transfer model TUV - version 4.1b (Total ultraviolet-visible model, *Madronich, 1993*) was employed for the calculation of the UV-B data correction factor. SZA dependant UV-B corrections to values corresponding to the same total ozone value were calculated using a discrete-ordinates radiative transfer scheme.

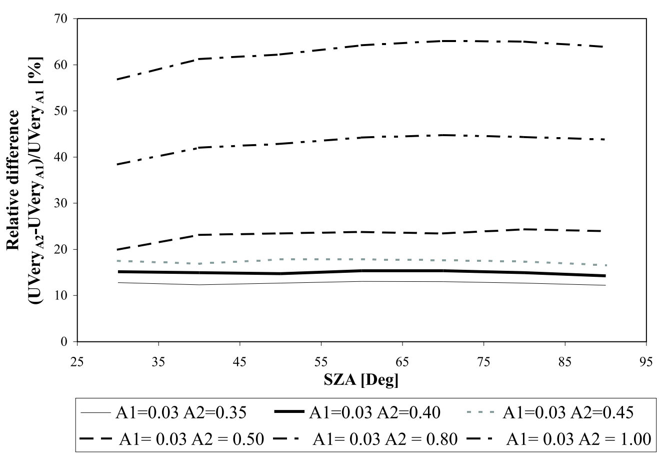

Fig. 3. Relative increase in erythemal irradiance induced by snow cover characterized with different snow effective albedo A_2 values in the range 0.35-1.00 as a function of SZA modeled by the TUV model at SL. Effective albedo of no-snow surface A_1 was 0.03. Discrete–ordinate radiative transfer scheme, total ozone equal 300 DU, aerosol optical depth (aerosol optical depth for the radiation with wavelength 340 nm AOD_{340}) $AOD_{340} = 0.2$, SSA = 0.95, AP = 0.66 and SL coordinates were applied in the radiative transfer model.

Total ozone was measured at Poprad-Gánovce (about 17 km far from Stará Lesná) by a Brewer spectrophotometer MKIV. The instrument is regularly calibrated by a world standard device. The small distance between the Poprad-Gánovce and investigated observatories enables us to consider the Poprad-Gánovce total ozone as representative for the High Tatras area. Linear interpolation of the total ozone daily courses obtained by the direct-sun method was used to get the total ozone value appropriate to the hourly UV-B irradiance.

As a cloud effect on observed irradiances is predominant, it was necessary to select clear-sky data. The sunshine duration registered continuously for the whole hour and average hourly cloudiness within 0/10 – 1/10 were the criteria for cloudless data selection. A requirement for the continuous hourly sunshine duration leads to the exclusion of data measured close to sunrise and sunset, and also to removing data differently affected by the height of the horizon at both stations. But there were still many situations, when the global irradiance at lower position exceeded the irradiance measured at higher altitude, especially during the cold half-year. The decrease in the global irradiances with increased altitude was also registered when no cloudiness was in the sky. This probably relates to some instrumental problems, since these situations occurred only at SZA $> 50°$ (Fig. 5B).

Both the investigated places are characterized by high average daily cloudiness – 7/10 and 6/10 for SP and SL respectively. That is the reason, why a number of the investigated cloudless data was relatively low for this study (Tab. 1). Strong convection activity is observed at the south slopes of the High Tatras. Cloud creation during warm half-year, immediately above SP, leads to insufficient clear-sky hourly radiation data observed by low SZA. In the warm half-year, at least 1/10 of cloud cover can manifest in the hourly radiative flux values.

An average number of days with snow is 206 at SP. Snow is there usually registered from October to May. The maximum snow-cover thickness is 46 cm in March. At SL, the average number of days with snow is 114. The maximal average snow thickness of 16 cm is registered in February. A continuous snow cover is there usually registered from December to March, in a few cases in April. The lasting of the snow cover at observatories determined the range of the SZA for snow condition to more than 40° at SP and up to 55° at SL.

3. Results and discussion

The investigation of snow and altitude effects (also with respect to the SZA) on solar radiation was limited to clear-sky conditions. Radiation data were subdivided into four groups with respect to snow cover at both places:

I. No snow

II. Continuous snow at both stations

III. Continuous snow cover at the higher altitude (SP), no snow at the lower position (SL)

IV. Incoherent, melting snow at the higher station (SP), no snow at the lower station (SL)

An averaged total ozone, surface reflectance R/G and the diffuse irradiance fraction in total irradiance D/G for SZA = 60° are summarized in Tab. 1 for

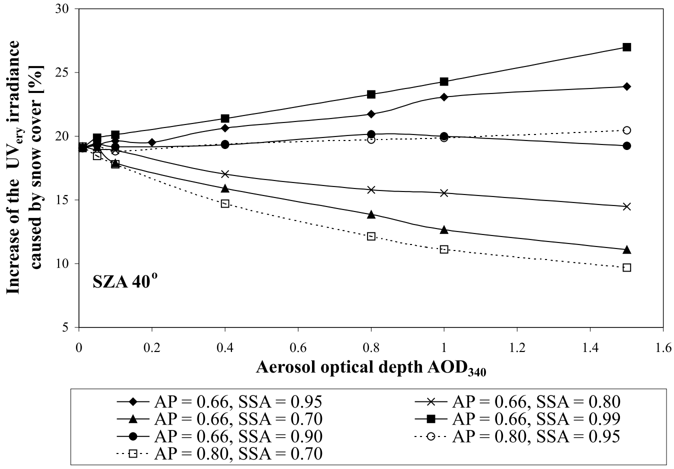

Fig. 4. Relative increase in the erythemal irradiance induced by snow cover as a function of aerosol optical depth (aerosol optical depth for the radiation with wavelength 340 nm AOD_{340}) modeled using the TUV model by SZA = 40°. Aerosol optical characteristics were expressed by typical range of SSA (0.70–0.99) and AP (0.66–0.8). Discrete–ordinate radiative transfer scheme, total ozone equal 300 DU, no-snow effective albedo equal 0.02, snow effective albedo equal 0.50, and SP coordinates were applied in the radiative transfer model.

Tab. 1. Number of data N, average surface reflectivity R/G, total ozone O_3 and contribution of the diffuse component to the total irradiance D/G summarized for data groups I - IV (see text for classification criteria). Number of the total irradiance data after removing cases, when the SL total irradiance exceeds the total irradiance at SP, is in brackets

Data	N		R/G [%]		O3 [DU]	D/G SZA=60° [%]	
group	UV-B	G	SP	SL		SP	SL
I	86	90(61)	18±2	25±2	290±31	11±5	14±5
II	147	147(77)	83±10	73±4	351±39	10±4	11±5
III	92	92(63)	71±12	23±1	303±122	11±2	13±3
IV	70	70(58)	30±22	24±1	327±35	13±4	17±2

every data group. The irradiances observed by continuous snow cover were compared with the irradiances observed without any snow at both observatories. They were plotted as a function of SZA and fitted by polynomial function to determine effect of snow on them (Fig. 1A–D).

Root mean square error (RMS) of UV-B data fit was in the range of 9–18%; RMS error for the global radiation was within interval 7–9%. The RMS error of 38% was calculated for the reflected radiation and the surface covered by snow at SP. As the diffuse radiation is very sensitive to atmospheric turbidity changes, the effect of snow is not perspicuously expressed in the data. The air turbidity variations cause too large a spread of the diffuse radiation data to determine change in this irradiance induced by snow.

The UV-B irradiance was recalculated to the equal total ozone value. The remaining spread in the data can be considered as a consequence of the surface reflectivity changes and the atmospheric turbidity variations. The higher RMS error obtained for the UV-B radiation, in comparison with the total radiation, indicates higher sensitivity of the UV-B data to the surface albedo and the turbidity variations.

The relative change in the total, UV-B, and reflected radiation caused by snow was determined comparing data groups I and II. Results are summarized in Tab. 2. The relative increase in irradiances due to snow is determined with large error, as low number of cloudless data was available during the investigated period, and also as a consequence of natural variability of the fitted data induced by the reflectivity and turbidity variations. An increase in differences between the snow and no-snow UV-B and global irradiances by large SZA can be explained by the non-Lambertian character of the snow surface.

Tab. 2. A relative increase in the UV-B, total global (G) and reflected (R) irradiances induced by snow cover in % at Stará Lesná (SL) and Skalnaté Pleso (SP) determined for selected solar zenith angles (SZA). Data categories II and I were compared. Values determined from all total irradiance data, included also cases when the total irradiance at SL exceeds values at SP, are in brackets

Locality	SZA	$(UVB_{snow}$-$UVB)/UVB$ [%]	$(R_{snow}$-$R)/R$ [%]	$(G_{snow}$-$G)/G$ [%]
SL	55°	15±46	253±28	12±3(12±2)
	60°	16±66	250±31	14±4(14±3)
	65°	13±100	249±37	18±4(17±3)
	70°	10±172	255 ±47	23±6 (21±5)
	75°	14±355	273±70	32±8 (27±7)
SP	55°	13±38	411±204	10±3 (8±2)
	60°	13±54	422±237	12±3 (12±3)
	65°	12±83	435±286	14±4(15±3)
	70°	14±149	455±369	19±5(19±4)
	75°	27±342	488±536	26±8 (24±6)

The relative increase in the UV-B irradiance caused by snow cover within 12–13% was determined at SP and within 13–16% at SL for the SZA in range 55° – 65°. The total irradiance increase reaches values of 10–14% at SP and 12–18% at SL (SZA in range 55° – 65°). Relative increase in total radiation by snow cover is greater for large SZA, but these values are also determined with high uncertainty. The relative increase in the total irradiance induced by snow did not change significantly after removing irrelevant cases, when global irradiance at SL exceeded values at SP. Small decrease in the snow effect on the UV-B and global irradiances at SP (in spite of higher snow reflectivity at SP) relates to a decrease in the turbidity and consequently to the decrease in the diffuse component of the total irradiance at higher altitude. Also no significant change in the ratios UV-B/G and D/G was assessed by snow surface in comparison to snow-less surface.

To demonstrate the effect of remote snow areas on investigated irradiances, the condition with continuous snow cover at SP and no snow at foothill position was studied separately (Fig. 2A-B). Increase in global irradiance of 7–10% was determined at SL as a consequence of snow presence at high altitudes (Fig. 2B). However, no significant increase in the UV-B irradiance was found in this case at SL. A relative increase in the UV-B irradiance of 4–7% for SZA within 55° – 65° and of 11–14% for SZA in range 40° – 50° was found for data group III in comparison with data group I at SP. Comparison of irradiances distributed to the data group II and III shows decrease in the snow effect on the UV-B and total irradiance of 2–7% for data group III at SP.

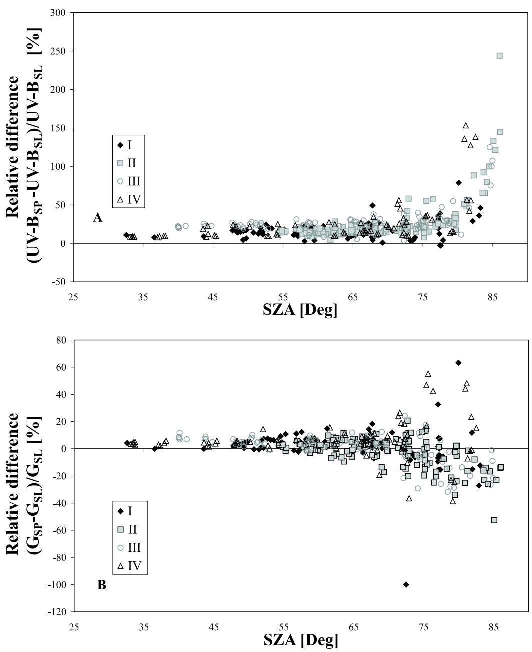

Fig. 5. A relative difference between the UV-B (A) and total (B) irradiances observed at SP and SL as a function of the SZA. Data are subdivided into categories I-IV.

Two days characterized by equal visibilities 100 km, 50 km, 50 km observed in the morning, noon and evening respectively, were selected to determine effect of snow reflectivity variation on the UV-B and global irradiances. The D/G ratios were also very close for both days – D/G = 0.10 on March 25, 2003 and D/G = 0.12 on April 15, 2003. The average morning reflectivity was 0.76 on March 25 and 0.66 on April 15 at SP. Hourly UV-B irradiances were corrected for total ozone 300 DU, which was extremely important, because total ozone varied more than 50 DU during April 15. A vertical change in the UV-B irradiance between SP and SL was 24% during both days. As both days belong to data group III (no significant change in the surface reflectivity was at SL - R/G = 0.23 during both days), higher global and UV-B irradiances were expected on March 25 at SP. Really, the morning UV-B irradiances were by 5–6% higher on March 25 at SP. The total irradiance changes less than 1%. However, increase in the UV-B irradiance of 6–10% was observed at this day at SL too. Whereas no reflectivity variations were detected at SL, we think, that the change in the UV-B irradiance was mostly induced by the turbidity variation also at SP. Also no significant difference between the morning and afternoon UV-B data was found on April 15 at SP, in spite of different surface reflectivities observed in the morning and in the afternoon hours. More exact knowledge on atmospheric turbidity than just the information on horizontal visibility is necessary to study the effect of snow reflectivity variation on the solar UV-B radiation, which is more sensitive to atmospheric turbidity change, than the radiation in the visible range of the solar spectrum.

The relative change in an erythemal UV radiation (spectrally weighted by the action spectrum for human erythema), as a result of different effective albedos, was simulated using the radiative transfer model TUV. Relative differences between erythemal irradiance observed by no-snow surface albedo equal to 0.03 and snow effective albedo within range 0.35–1.00 are shown in Fig. 3. The aerosol optical depth for radiation with wavelength 340 nm (AOD_{340}) equal to 0.2 was appointed to the air turbidity characterization. The single scatter albedo (SSA) and asymmetry parameter (AP) values were selected to represent clean continental air (*Wenny et al., 1998*). Additional input parameters are presented in the figure. The snow induced increase in the UV-B irradiance in range 13–16% determined from measurements corresponds to the snow effective albedo values within interval 0.35–0.40 at SL. Similar values of effective albedo of 0.3–0.4 were found by *Pachard et al. (1999)* in the French Alps. *Blumthaler and Ambach (1988)* determined significantly higher effective albedo of fresh snow of 0.95 for the solar erythemal radiation. Such high effective albedo values of 0.7–0.8 were measured in winter at the Jungfraujoch, Switzerland (*Blumthaler and Haferl, 1999*).

The effective albedo of snow decreased to value of 0.45 in spring at this place.

Fig. 4 documents the effect of different aerosol optical properties on the erythemal irradiance variability by the snow condition. The aerosol optical parameters were simulated in typical range of the SSA and AP (*Madronich, 1993a*) using SP coordinates.

The vertical change in the studied irradiances was determined with respect to snow presence at compared places (Tab. 3). Whereas simultaneously measured irradiances were compared, no corrections were applied to the data. However, the comparison of irradiances measured with different instruments leads to the increase in the differences between the measurements by the large SZA due to different instrumental characteristics (relative angular response, relative spectral response) – Fig. 5A-B.

The irradiance measured at SL was plotted versus the irradiance at SP to determine the mean value of its vertical gradient. The difference between the linear data fit slope and the slope equal to 1 represents a relative decrease or increase in the investigated irradiances between SL and SP observatories (Fig. 6).

Tab. 3. Relative change in the UV-B, total (G), reflected (R) and scattered (D) irradiances between SP and SL in %. A negative value means the decrease in an investigated irradiance with increasing altitude and vice versa. Vertical change in the ratios UV-B/G and D/G is presented, too. Values determined from all total irradiance data, included also cases when the total irradiance at SL exceeds values at SP, are in brackets

Data group	UVB [%]	G [%]	D [%]	R [%]	D/G [%]	UVB/G [%]
I	15.2±0.4	5.8±0.4 (3.4)	-24±1	-24.4±0.5	-24±1	8.9±0.4
II	16.7±0.4	4.6±0.3(0.0)	-8±2	14±1	-10±2	12.1±0.4
III	23.1±0.4	6.2±0.4(3.9)	-9±2	331±5	-17±2	16.8±0.4
IV	14.5±0.7	6.3±0.5(3.8)	-12±1	9.5±0.7	21±1	9.5±0.6

The vertical change in the investigated irradiances is summarized in Tab. 3. The UV-B irradiance increase of 15% /1000 m was determined by no-snow condition. The total irradiance vertical increase (after removing cases, when SL global irradiances exceed the SP values) was only 6% in this case. Higher vertical change in the UV-B irradiance relates to major attenuation of shorter wavelengths by atmospheric aerosol. Similar results were obtained by *Blumthaler et al. (1997)* in the Alps. They found an increase in the erythemal irradiance of 18% /1000 m and an increase in total irradiance of 8% /1000 m. *Zarrati et al. (2003)* found less increase in erythemal irradiance of 6% /1000 m in the Bolivia. But these measurements were performed at very high altitudes (3420–5200 m a.s.l.) and these places can be probably noted for very low atmospheric turbidity.

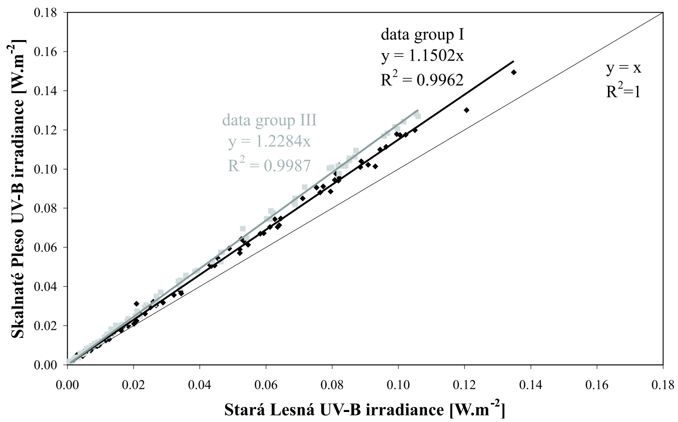

Fig. 6. The UV-B irradiance at SL versus the UV-B irradiance at SP. The linear least square data fits together with regression equations y = k.x and squared correlation coefficient R^2 are shown for data groups I (black color) and III (gray color), in comparison with linear function y = x with slope k = 1. Vertical change of the UV-B irradiance between SP and SL 15% and 23% was found for data groups I and III, respectively.

The combined effect of increased altitude and surface albedo induced large vertical gradient of 23% /1000 m for data group III (Fig. 6). Also the contribution of the UV-B irradiance to the total global irradiance increases with altitude. The increase in the ratio UV-B/G was 9% /1000 m and 17% /1000 m for data groups I and III, respectively. The diffuse irradiance and its contribution to the total radiation decreased with altitude for all data groups. However, the decrease in the ratio D/G was partly compensated by higher reflectivity at SP by the data groups II-IV.

Two days were selected from data group I to demonstrate the effect of different turbidities on the vertical gradient of the investigated radiative fluxes. On September 9, 2003, horizontal visibility decreased from morning distance 100 km to noon value 60 km. The morning and noon visibilities were equal 100 km on July 4, 2004 at SP. Decrease of the ratio D/G with height was 31% /1000 m on September 9 and even the D/G increase of 3% /1000 m was determined on July 4. The vertical gradient of total irradiance was 5% and 6% for the days with lower and higher visibility, respectively. The UV-B irradiance increases 19% /1000 m and 10% /1000 m were determined for high and low aerosol content, respectively. Also the vertical gradient of the ratio UV-B/G increased from 3% /1000 m (low turbidity) to 14% /1000 m (higher turbidity).

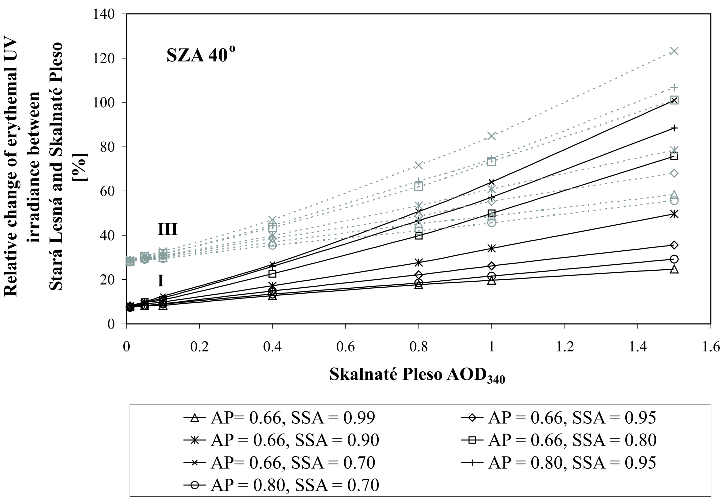

Fig. 7. Relative change in the erythemal irradiance between SL and SP as a function of air turbidity characterized by aerosol optical depth AOD_{340} (aerosol optical depth for the radiation with wavelength 340 nm) for data group I and III modeled by TUV model. An aerosol was characterized by usual range of SSA (0.70–0.99) and AP (0.66–0.80). Snow surface was characterized by effective albedo of 0.5, snow-less surface effective albedo of 0.02 was used in the model. Discrete-ordinate radiative transfer scheme, total ozone 300 DU, SP and SL coordinates were applied in the TUV model.

The importance of the air turbidity variation effect on the vertical change in the UV-B radiation is also demonstrated by the TUV model simulation for data groups I and III (Fig. 7). The variability of the aerosol optical characteristics (SSA, AP) affects the vertical gradient of the UV-B irradiances significantly by high atmospheric turbidity values.

4. Conclusions

The global UV-B irradiance variability was studied at two sites in the High Tatras under cloudless condition. The change in the UV-B irradiance caused by the surface albedo variation and altitude increase was compared with the total irradiance sensitivity to the noted phenomena.

The increase in the surface reflectivity caused by snow cover leads to an increase in the UV-B irradiances of 13–16% by SZA 55° – 65°. There were no significant differences between SL and SP observatories. Higher reflectivity of snow surface and less contribution of diffuse component to total irradiance probably compensate snow induced increase in the UV-B irradiance at SP in comparison with SL. It leads to nearly equal snow effect on the UV-B radiation at SP and SL. For the total irradiance, snow induced increase of 10–14% was determined for the SZA in range 55° – 60°. The snow induced increase in the total radiation by 2–7% lower at SP than at SL.

The determined increase in the UV-B irradiances caused by snow surface corresponds to the modeled effective albedo within range 0.35–0.40. The dependence of snow effect on the SZA is higher by measured UV-B and total irradiance data than by modeled situations.

Uncertainties of the relative differences between the snow and no-snow irradiances are large partially due to low number of the clear-sky data and partially due to natural variability of irradiances, especially UV-B, caused by air turbidity and surface reflectivity changes.

The comparison of irradiances split into data groups II and III shows a decrease in the snow effect on the UV-B and the total irradiance for data group III of 2–7% at SP. It relates to the decrease in the snow reflectivity at SP and also to less amount of the diffuse radiation originated by reflection on the snow-less areas below SP during spring months. No significant change in the UV-B irradiance was found comparing the data group I (no snow) and the data of group III (snow at remote mountain slopes) at SL. But total irradiance was by 7–10% higher during spring in comparison with no-snow summer data at SL.

The vertical gradient of the UV-B radiation of 15%–17% /1000 m is significantly higher than the vertical change in the total radiation (5–6% /1000 m) for equal snow or no-snow surfaces at compared places. It relates to more pronounced attenuation of radiation with short wavelengths by atmospheric aerosol. In spring months, when continuous snow persists at high altitudes, vertical gradient of the UV-B irradiances increased to value of 23% /1000 m. Since the vertical change in total irradiance is lower than the vertical gradient in the UV-B radiation, also the UV-B contribution to the total irradiance increases with altitude. The relative change in the UV-B/G ratio of 9% and 17% was determined between SP and SL for data groups I and III, respectively.

The TUV model simulations document the importance of exact knowledge on aerosol content and its properties in the UV-B irradiance variation study under clear-sky condition.

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