Variability in snow depth time series in the Adige catchment

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ABSTRACT

Study region: The Upper and Middle Adige catchment, Trentino-South Tyrol, Italy.
Study focus: We provide evidence of changes in mean seasonal snow depth and snow cover duration in the region occurred in the period from 1980 to 2009.
New hydrological insights for the region: Stations located above and below 1650 m a.s.l. show different dynamics, with the latter experiencing in the last decades a larger reduction of average snow depth and snow cover duration, than the former. Wavelet analyses show that snow dynamics change with elevation and correlate differently with climatic indices at multiple temporal scales. We also observe that starting from the late 1980s snow cover duration and mean seasonal snow depth are below the average in the study area. We also identify an elevation dependent correlation with the temperature. Moreover, correlation with the Mediterranean Oscillation Index and with the North Atlantic Oscillation Index is identified.

1. Introduction

Seasonal snow cover duration and mean seasonal snow depth have received significant attention in studies dealing with climate change (Gobiet et al., 2014; Beniston and Stoffel, 2014), hydrology (Barnett et al., 2005; Berghuijs et al., 2014; Schneeberger et al., 2015), nitrogen and carbon cycle (Williams et al., 1998; Euskirchen et al., 2006), ecosystem functioning (Keller et al., 2005), soil processes (Groffman et al., 2001) and economy (Steiger and Stötter, 2013; Beniston, 2012a). Most of the existing studies focus on the Swiss and Austrian Alps (e.g. Beniston, 1997; Hantel et al., 2000; Schöner et al., 2009; Marty, 2008; Laternser and Schneebeli, 2003; Beniston et al., 2003). Much less attention has been devoted to the Southeastern Alps, in the Italian territory (Bocchiola and Rosso, 2007; Valt and Cianfarra, 2010; Terzago et al., 2010; Pistocchi, 2016), despite the importance of snow dynamics in timing water resources availability (Chiogna et al., 2016; Callegari et al., 2015), streamflow generation (Penna et al., 2013, 2017; Chiogna et al., 2014; Engel et al., 2016) and ecosystem functioning (Lencioni et al., 2011; Esposito et al., 2016).

In this paper, we analyze the dataset available covering the Adige catchment and five surrounding stations highly correlated with those within this river basin. We focus on the period of 1980–2009, which is the richest of data for the area investigated. Furthermore, the analysis of snow dynamics in this time frame is of particular interest, since the 1990s have been an unusually warm and dry period for this region (Brunetti et al., 2009). Deepening the knowledge of snow dynamics South of the main Alpine ridge is relevant because it is influenced by a meteorological circulation pattern that differs from the territories in the northern side of the mountain chain (Xoplaki et al., 2004; Buzzi and Tibaldi, 1978). Owing to the different circulation patterns, different snow dynamics may be observed in the two regions, and the validity of the outcomes obtained for the Swiss and Austrian Alps does not necessarily apply also to the Italian Alps. Moreover, the Adige catchment can be assumed as representative of the Southern Alpine ridge.

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First, we study the difference of the behavior of stations above and below 1650 m a.s.l. and the changes in mean seasonal snow depth and snow cover duration occurred after 1987, showing that temperature changes may explain the observed snow dynamics, in particular for low elevation sites. Then, we perform a wavelet analysis (Torrence and Compo, 1998) of the mean seasonal snow depth and of the snow cover duration at four elevation classes (below 1350 m a.s.l., between 1350 m a.s.l. and 1650 m a.s.l., between 1650 m a.s.l. and 2000 m a.s.l. and above 2000 m a.s.l.). This allows us to identify changes occurring in the snow depth and snow cover duration signals at various temporal scales. Finally, we apply the wavelet coherence analysis (Grinsted et al., 2004) in order to investigate the relationship between the variations of the mean seasonal snow depth and that of climate indices such as the North Atlantic Oscillation Index (Hurrell, 1995) and the Mediterranean Oscillation Index (Palutikof, 2003). Several studies tried to link the North Atlantic Oscillation Index (NAOI) to the snow variability in order to explain it, however these results are not univocal (Scherrer and Appenzeller, 2006; Schöner et al., 2009; Scherrer et al., 2004; Durand et al., 2009; Beniston, 1997).

Section 2 describes the study area and the dataset available for this study, Section 3 is devoted to the description of the results, which are then discussed in Section 4, and finally Section 5 summarizes the conclusions.

2. Material and methods

2.1. Study area

The Adige catchment is one of the most important river basins in Italy, mainly for its large catchment area (12,100 km^2) and length (409 km) and for its hydropower generation. A recent review of Chiogna et al. (2016), described the hydrological conditions in the catchment and its chemical and ecological status in details. The most important tributaries of the Adige catchment are located in the Alpine part of the basin, and hence they are strongly influenced by snow dynamics (Mei et al., 2014; Tuo et al., 2016). An increasing concern is rising due to the effects of climate change in this area, since this has already shown important implications for water resources management, and above all, for hydropower production and for winter tourism (Majone et al., 2016). In terms of atmospheric circulation, the Adige catchment is mainly affected by southwest weather patterns and lee cyclones (Xoplaki et al., 2004; Buzzi and Tibaldi, 1978). Brunetti et al. (2006a, 2009) observed a slight negative trend in precipitation for the South Western area of the Greater Alpine Region. The Adige river basin belongs to this area, however, more recent studies, specifically addressing precipitation trends in the river catchment (Lutz et al., 2016; Adler et al., 2015) did not identify significant trends. Moreover, Adler et al. (2015) indicates two additional drawbacks of the precipitation dataset available for the study area. The first is that very few stations are available for elevations above 2000 m a.s.l. The second is the significant underestimation of winter precipitation due to undercatch of the rain gauges during snowstorms. Therefore, in this study we will not investigate the correlation between snow and the available precipitation dataset. Brunetti et al. (2006a,2009 observed a positive and statistically significant temperature trend in the river basin. Moreover, the quality of the available time series was verified (see e.g. Adler et al., 2015). Therefore, we focus our analysis on the correlation between temperature and snow.

2.2. Description of the dataset

Snow depth time series of meteorological stations located in the Adige catchment are a relevant source of information to study snow dynamics in the Alpine region (Fig. 1) because of their spatial distribution over a wide elevation range and time spanning. In order to extend the dataset as much as possible we included five additional stations (Malga Bissina, Caoria, Brocon – Marande, San Martino di Castrozza and Sexten), which are outside the catchment, but very close to its boundaries. This choice was justified by the high correlation ($r > 0.9$) of the time series at similar elevations in this region. The dataset is therefore composed by 37 stations. The stations have been grouped into four elevation classes, as shown in Table S1: below 1350 m a.s.l. (14 sites), between 1350 m and 1650 m a.s.l. (12 sites), between 1650 m and 2000 m a.s.l. (7 sites) and above 2000 m a.s.l. (4 sites).

The dataset consists of daily data for the period 1st November–30th April between two successive years of the time series. Conventionally, we attributed this period to the starting year, e.g. the season 1990 is intended to be comprised between 1st November 1990 and 30th April 1991. The time series may be formed by merging data obtained at stations which may have been relocated during the period from 1980 to 2009. Moreover, the time series are formed merging data from three different sources, according to quality indices criteria (Marcolini et al., 2017). Before the merging operation, we have checked the quality of the data. The highest quality index was assigned to manual data (measured from operators directly in the field), the second highest to automatic data (measured from automatic instruments), while historical data (collected from different sources, such as the Zentralanstalt für Meteorologie und Geodynamik (ZAMG) of Vienna and the Hydrographic office of the Province of Trento) are considered the least reliable, because measuring procedures and station locations are not always well documented. The merging of the data was performed to obtain single and longer time series for each site. Short gaps in the time series (i.e., shorter than 14 days), were filled by support vector machine regression (Smola and Schölkopf, 2004) performed by applying the Matlab toolbox Spider (http://www.kyb.tuebingen.mpg.de/bs/people/spider/), which uses the snow depth of the two best correlated stations and, if available, snowfall, temperature and precipitation data of the examined stations, as input variables for the regression.

Verifying the homogeneity of the time series is an important prerequisite for detecting trends and investigate climatic changes. Therefore, breakpoints in the time series (Auer et al., 2007; Brunetti et al., 2006a) caused by station relocation or merging of different sources should be detected before carrying on further analyses. In this article, we only consider homogenous time series in the period from 1980 to 2009. To check for homogeneity of available snow depth time series, we applied the Standard Normal Homogeneity Test (SNHT) (Alexandersson and Moberg, 1997; Alexandersson, 1986; Marcolini et al., 2017).
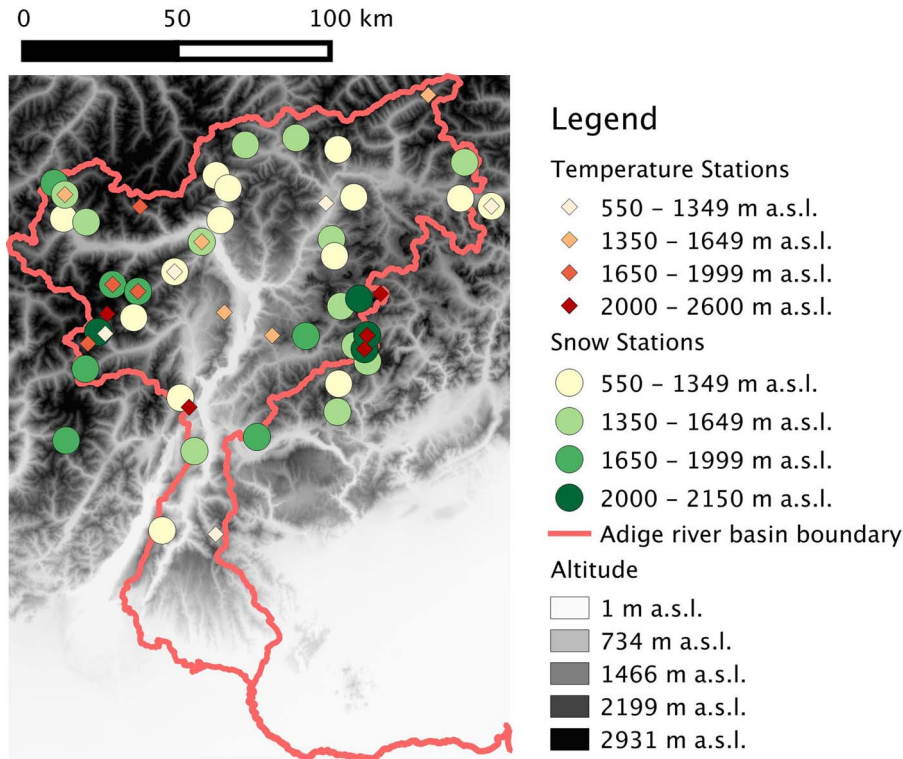
Fig. 1. Location of the analyzed time series.

Temperature data were also collected from available databases of the meteorologic agencies of the Autonomous Province of Trento (http://www.meteotrentino.it) and the Autonomous Province of Bolzano (http://www.provincia.bz.it/meteo). The stations we considered in this manuscript are summarized in Table S2.

In this study, we aim at investigating also the correlation between snow dynamics and climatic indices, such as the NAOI and the MOI. For both climatic indices, we considered the mean value computed over the period from 1st November of a given year to 30th April of the following year. The NAOI is defined as the normalized pressure difference between Stykkisholmur (Iceland) and Lisbon (Portugal) (Hurrell, 1995). In the southern part of Europe, and in particular in the Alpine region, the precipitations are negatively correlated with the NAOI. On the contrary, a positive correlation has been observed between the NAOI and the Alpine temperature (Beniston, 2012b). The MOI is defined as the normalized pressure difference between the Gibiltrar's Norther Frontier and Lod Airport in Israel (Palutikof, 2003).

2.3. Investigated variables

Snow cover duration (SCD) was defined as the total number of days in a given snow season (from 1st November of a given year to 30th April of the following year) with snow depth higher than 30 cm (Durand et al., 2009; Valt and Cianfarra, 2010). This threshold was selected because of its practical implication for winter tourism (Marty, 2008). The mean seasonal snow depth HS was computed by averaging the daily snow depth between 1st November of a given year and 30th April of the following year.

Another quantity of interest we investigated in the present study was the relative fluctuation of HS with respect to its mean $\langle HS \rangle$

$$\mathrm{HS}_m = \frac{\mathrm{HS} - \langle \mathrm{HS} \rangle}{\langle \mathrm{HS} \rangle}. \tag{1}$$

The use of a standardized variable allows to compare the results of data collected at different elevations. Similarly we defined the fluctuation of SCD as

$$\mathrm{SCD}_m = \frac{\mathrm{SCD} - \langle \mathrm{SCD} \rangle}{\langle \mathrm{SCD} \rangle}. \tag{2}$$

2.4. Wavelet analysis

Wavelet analysis has been shown to be a useful tool for detecting localized variations in the modes of variability within time series (e.g. Lau and Weng, 1995; Coulibaly and Burn, 2004; Zolezzi et al., 2009; Guan et al., 2011; Carey et al., 2013). The wavelet analysis

is performed by decomposing the time series into a transformed variable in time-frequency space to determine both the dominant modes of variability and how these modes vary in time. We performed wavelet analysis by using the Matlab toolbox developed by Torrence and Compo (1998).

The continuous wavelet transform looks for similarity between a signal and a well-known mathematical function (Labat, 2005). The major difference to the Fourier transform, which is also the strength of this kind of analysis, is that the well-known mathematical function, which for the wavelet analysis is a wavelet function, is applied several times with different scales to the analyzed time series and at different temporal positions. Wavelet transform allows us to determine not only the frequency content of a signal, as also the Fourier analysis can do, but also the frequency time-dependence (Labat, 2005).

Similarly to other studies analyzing variability in climate and hydrological signals (see e.g. Lau and Weng, 1995; Coulibaly and Burn, 2004; Carey et al., 2013; Guan et al., 2011), the Morlet function was used as wavelet function for its ability to evidence fluctuations in time series.

The wavelet power spectrum (WPS), represents the energy of the scale *s* and is useful for the identification of fluctuation scales with the largest influence on the signal. In particular, a larger positive amplitude implies a higher positive correlation between the signal and the wavelet and a large negative amplitude implies a high negative correlation.

The wavelet transform assumes that the time series is periodic. Since hydrological time series are not periodic and only a fraction of the entire time series is available, a bias is introduced at the beginning and at the end of the time series. To remove this bias, Torrence and Compo (1998) suggested to pad the end of the time series with zeros. The introduction of the zero padding (not shown in the pictures) leads to the definition of a cone of influence where edge effects are significant and hence the results are uncertain (Torrence and Compo, 1998). The cone of influence is indicated by shadowing the area of the contour plot showing the wavelet transform influenced by edge effects.

In our article, in particular, we considered the analysis of both the global wavelet power spectrum and the signal integrated from 2 to 8 year scales. This allowed us to focus on the scales which lay within the cone of influence (i.e., results can be considered statistically significant), while filtering out the natural strong yearly periodicity of the snow signal.

Wavelet coherence (WTC) can be used to identify the scales where two time series *X* and *Y* focus on the power and how this may change in time (Torrence and Compo, 1998; Grinsted et al., 2004).

Wavelet coherence can be interpreted as the generalization in the scale-time space of the squared cross-correlation coefficient r^2 of two signals. Two time series display high coherence values (close to 1) for the time windows in which they are highly correlated. Similar to the classical correlation coefficient, when two time series display low coherence, the values are close to 0. The advantage of wavelet coherence analysis with respect to classical correlation analysis is the possibility to identify separately for each scale the coherence between two signals and eventually its variability over time. This allows us to identify at which scales and for which time windows two signals are more correlated.

For more details about wavelet transform and wavelet coherence the reader is referred to Torrence and Compo (1998), Grinsted et al. (2004), Labat (2005) and Appendix A.

3. Results

3.1. Statistical analysis

Fig. 2a and b shows an Hovmöller-like plot of the mean seasonal snow depth and of the snow cover duration at different elevations. The diagrams display the two variables, using a color code, as a function of time and elevation which are reported in the abscissa and the ordinate, respectively. A sharp change in the color pattern occurs at about 1650 m a.s.l. Notice that the color map of the Hovmöller-like plot of the snow cover duration has two different colors at a threshold of 100 days with more than 30 cm of snow depth during the winter season, since this is the level that is suggested to consider a site as economically profitable for winter activities (Valt and Cianfarra, 2010). It can be observed that sites below 1650 m a.s.l. rarely reach this threshold value (7% of the seasons for the stations below 1350 m a.s.l. and 17% of the seasons for the stations between 1350 m and 1650 m a.s.l.).

Considering the temporal variability of snow depth (Fig. 2a), we identify three periods. The first one ranges from 1980 to 1987, the year in which Marty (2008) identified a regime shift (based on the sequential *t*-test analysis of regime shifts) in the snow cover duration in Switzerland. The second one (from 1988 to 1999) and the third one (from 2000 to 2009) are identified observing the increase of the mean seasonal snow depth in Fig. 2a starting from 2000. To verify the statistical significance of our considerations both Kolmogorov–Smirnoff and Mann–Whitney tests have been conducted on mean seasonal snow depth data with a significance level of 5%. The results of the tests show that the period from 1988 to 1999 is characterized by significantly lower mean seasonal snow depth than the period from 1980 to 1987 (see Table 1). On the contrary, in the period from 2000 to 2009, the increase of mean seasonal snow depth was not identified as statistically significant with respect to the period from 1988 to 1999 (see Table 1). The same analysis performed on snow cover duration time series identifies a statistically significant change of the two lowest elevation classes (below 1650 m a.s.l.) and with 10% significance level (not shown).

Mean seasonal snow depth and snow cover duration have been analyzed considering also HS$_m$ and SCD$_m$ (Fig. 2c and d), and HS and SCD (Fig. 2e and f), aggregated according to the four elevation classes and smoothed using a 5-year moving average to eliminate short term fluctuations. The choice of a 5-year moving average is appropriate considering the length of 30 years of the investigated period and the length of the time intervals with low and high snow depth. The analysis of both mean seasonal snow depth and snow cover duration provided similar results. In Fig. 2c and d, it can be observed that stations located below 1650 m a.s.l. are affected by a larger variability than stations above 1650 m a.s.l. In particular, in the period from 1988 to 2005 fluctuations with respect to the
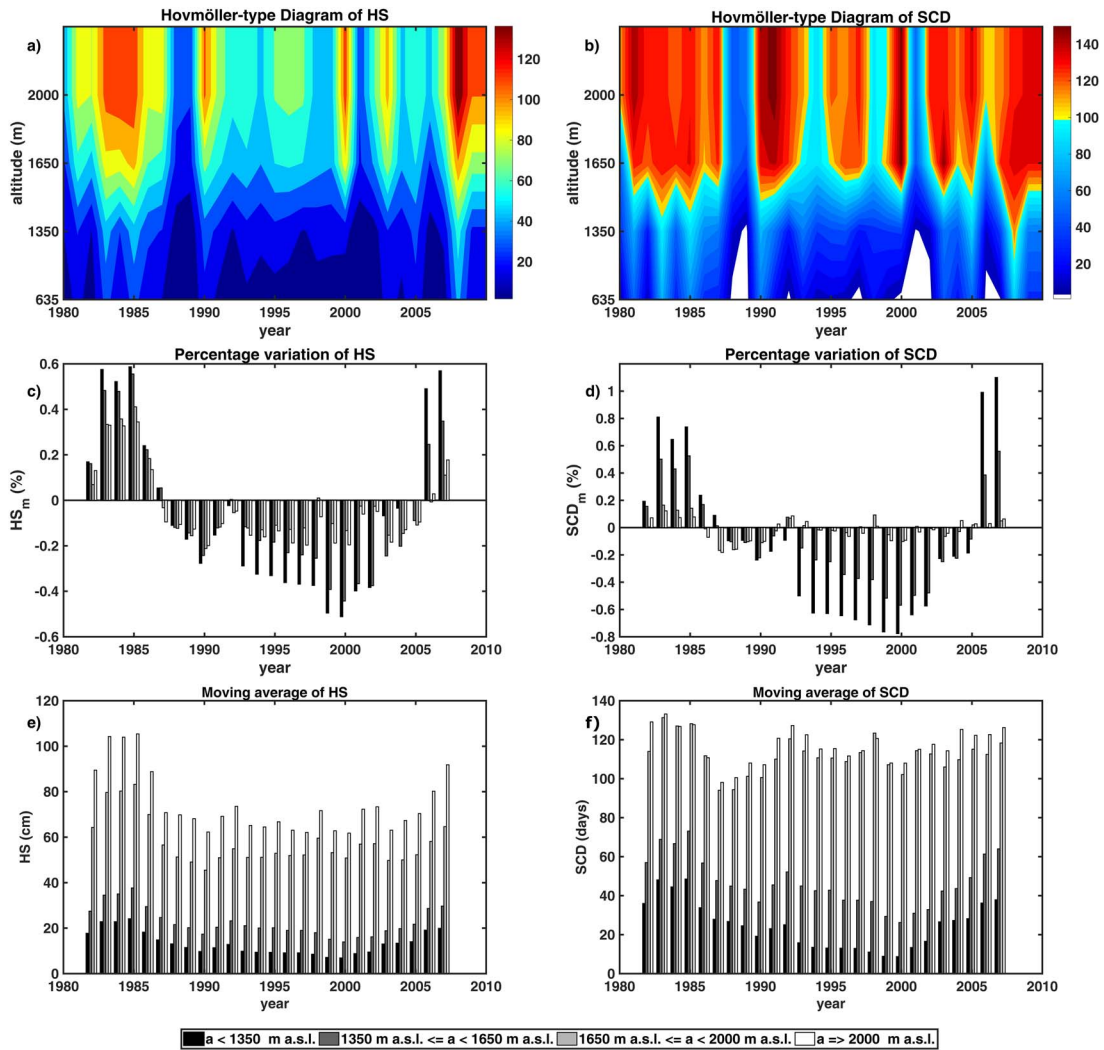
Fig. 2. Hovmöller-like diagram of mean seasonal snow depth (a), and snow cover duration (b), 5-year moving average HS$_m$ (c), and SCD$_m$ (d) as defined in Eqs. (1) and (2), respectively, and 5-year moving average of mean seasonal snow depth (e) and snow cover duration (f). The units of the colorbar of plot (a) is centimeters. The colourbar of the plot (b), instead, indicates the number of days with snow depth larger than 30 cm.

Table 1
P-values of the Kolmogorov–Smirnoff and of the Mann–Whitney tests applied to the average mean seasonal snow depth at different altitudes in three different periods: P1 stands for the period from 1980 to 1987, P2 stands for the period from 1988 to 1999, P3 stands for the period from 2000 to 2009.

Altitude class	Kolmogorov–Smirnoff		Mann–Whitney	
	P1–P2	P2–P3	P1–P2	P2–P3
$a < 1350$	0.01	0.23	0.03	0.63
$1350 \leq a < 1650$	0.01	0.56	0.03	0.63
$1650 \leq a < 2000$	0.01	0.27	0.03	0.35
$2000 \leq a$	0.00	0.23	0.01	0.25

mean, as defined by Eqs. (1) and (2), are generally more negative below 1650 m a.s.l. Furthermore, positive fluctuations occurring between 1980 and 1987 and between 2006 and 2009 are generally larger at lower elevations. Fig. 2d and f shows that fluctuations of snow cover duration for the two highest elevation classes are very low compared to those observed for the lowest elevation classes.

3.2. Correlation with temperature

Temperature exerts a strong influence on both mean seasonal snow depth and snow cover duration. The 5-year moving average of
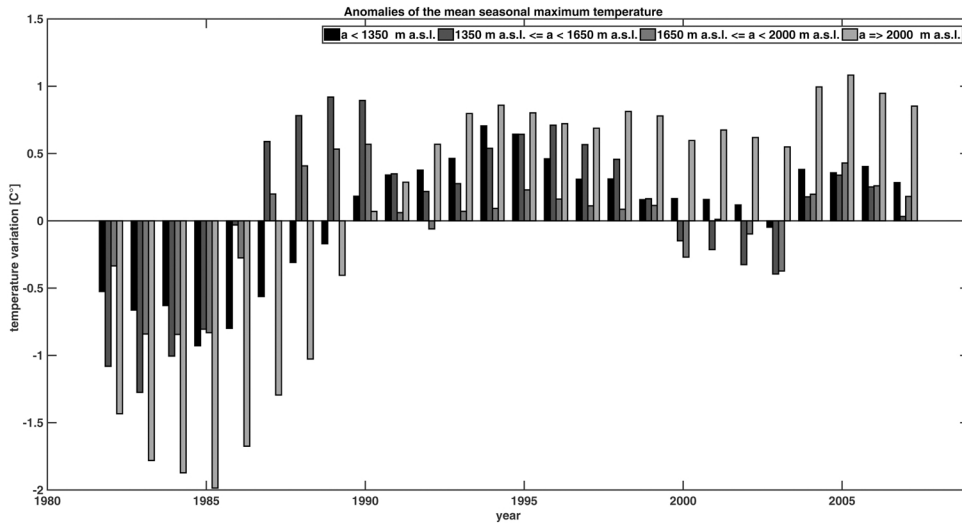
Fig. 3. Anomalies of mean seasonal maximum temperature at different altitude classes with respect to the average computed between 1980 and 2009.

the daily maximum temperature averaged over the winter season (from 1st November to 30th April) is shown in Fig. S1 in the supplementary material. The maximum temperature increases in the period from 1985 to 1990. Anomalies with respect to the mean of the maximum temperature of the winter season in the period from 1980 to 2009 are shown in Fig. 3. This highlights more clearly positive anomalies in the maximum winter temperature of the 1990s, which attenuate around 2000, though not at high elevations, to increase again after 2003. At high elevations (above 2000 m a.s.l.) anomalies are always positive after 1988 and the largest anomalies of all observation period are in the last years of the time series. Fig. 4 shows that mean seasonal snow depth and snow cover duration are negatively correlated with the seasonal mean of the maximum daily temperature. Table S3 in the supplementary material reports the results for the correlation analysis performed both using Kendall's tau τ_K (Embrechts et al., 2001; Nelsen, 2003) as well as Spearman's rho ρ_S (Embrechts et al., 2002) correlation coefficients. These are correlation coefficients based on the ranks of the data of the compared time series.
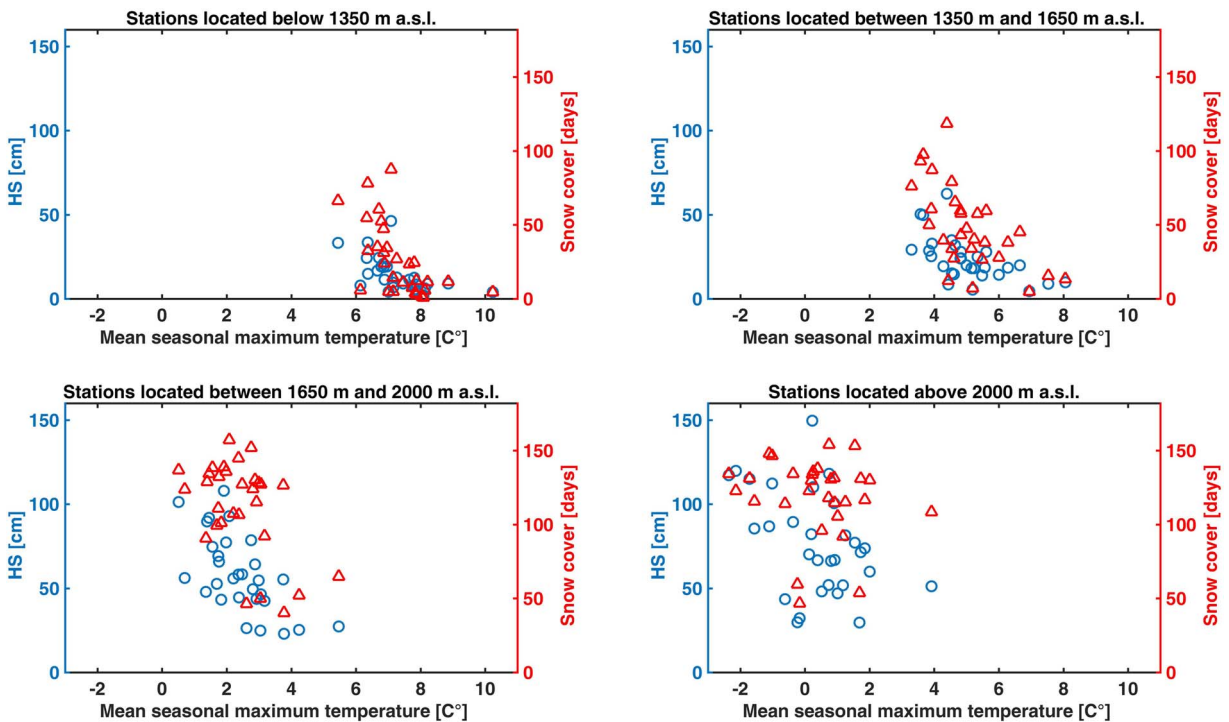

Fig. 4. Scatterplots between mean seasonal maximum temperature and mean seasonal snow depth at different altitude (circles referring to the left *y*-axis) and the snow cover duration (triangles referring to the right *y*-axis). Linear regression lines with 95% confidence bounds are shown in Fig. S5.
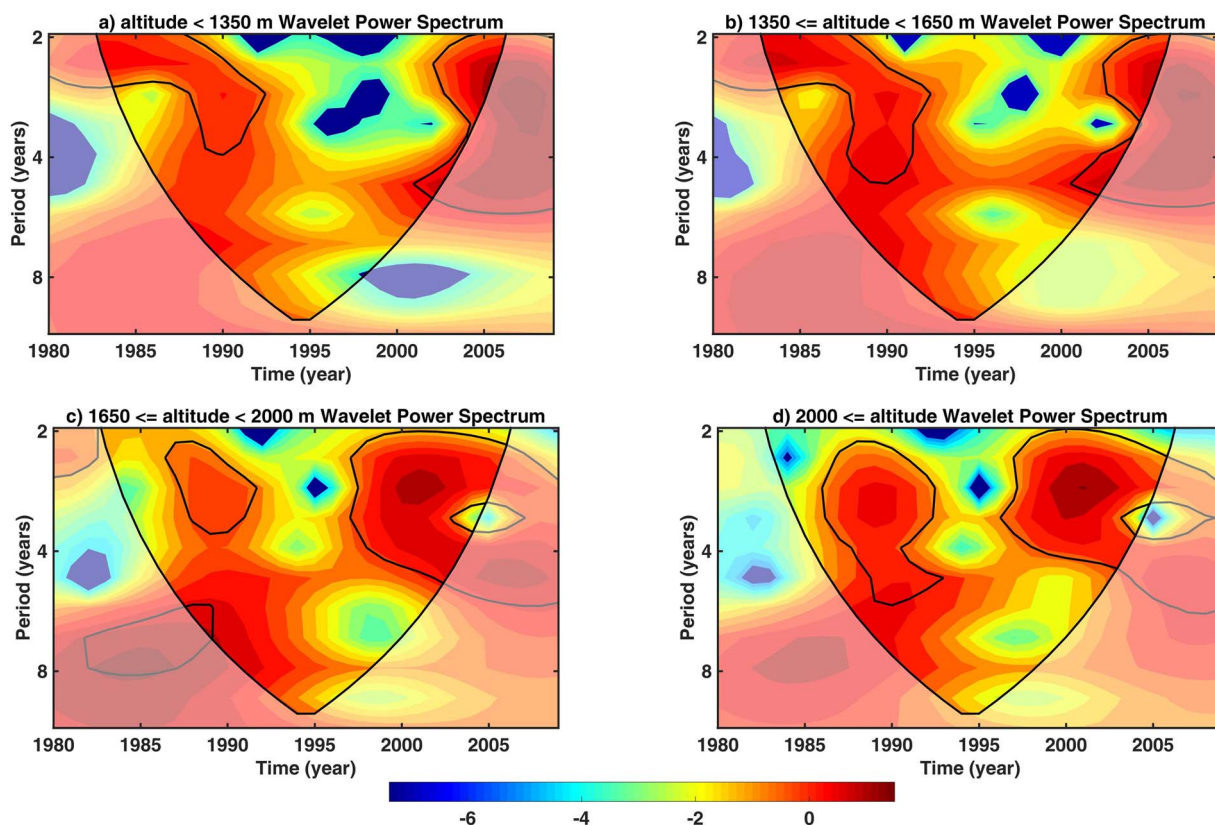
Fig. 5. Wavelet spectrum of the average mean seasonal snow depth for different altitude classes.

The negative correlation between temperature and HS is strongest for the stations located in the lowest elevation range ($\tau_K = -0.52$ and $\rho_S = -0.71$) and decreases with increasing elevation ($\tau_K = -0.30$ and $\rho_S = -0.42$ for the highest stations). The decrease in the values of the correlation coefficients with the altitude of the stations is even more evident when looking at the snow cover duration, for which we observe values varying from -0.54 to -0.15 for τ_K and values varying from -0.71 to -0.22 for ρ_S.

We also performed an hypothesis test of association with a confidence level of 95% based on the ρ_S and τ_K in order to verify the null hypothesis that two variables are not correlated (Hollander and Wolfe, 1973; Best and Roberts, 1975). The results are shown in Table S4 and confirm in general the statistical significance of the correlation.

3.3. Wavelet analysis

We analyzed the average of the homogeneous snow depth time series computed for the four elevation classes defined above. Fig. 5 shows the wavelet spectrum of the four time series pertaining to each elevation band between the periods of 2 and 10 years to focus on the characteristics of the signal beyond annual periodicity. The black contours are the 5% significance regions, using a red-noise background spectrum (Torrence and Compo, 1998). Red represents the highest power, while blue represents the lowest power and areas affected by edge effects, outside the cone of influence, are shaded. Low elevation stations, i.e. below 1650 m a.s.l., show high power at the period of two years from 1980 to 1992, with some spreading to the period of four years during the last 5 years of this interval (Fig. 5a and b). The power reduces after 1992, to increase again after 2001. The average behavior for stations above 1650 m a.s.l. (Fig. 5c and d) show high power at the 2 and 4 year periods only in the interval from 1987 to 1992 then it decreases, to increase again in 1996, hence earlier than for stations at low elevations.

Fig. 6 shows the wavelet power spectrum for the snow cover time series analysis. While for the two classes representative for low elevation sites (below 1650 m a.s.l.) the outcomes are similar to those obtained for the mean seasonal snow depth, a difference can be appreciated for the two elevation classes representative of high elevation sites (above 1650 m a.s.l.). For the highest elevation classes, the high power region between the 2 and 4 year periods is almost continuous between 1984 and 2005.

Figs. 7 and 8 shows the average of the power spectrum between 2 and 8 year periods for all available time series for both mean seasonal snow depth and snow cover duration, respectively. Figs. 7 and 8 represent the variance of the signal components comprised within the range between 2 and 8 year periods (Torrence and Compo, 1998). Low elevation snow depth time series, i.e., below 1650 m a.s.l., show a first peak in the variance between 1984 and 1989 followed by a reduction, which leads to a minimum between 1995 and 1999, followed by a strong rise in the variance, which has a peak around 2008. High elevation snow depth time series, i.e., above 1650 m a.s.l., do not always show a clear peak at the end of the 1980s, but they all show a period with lower variance during
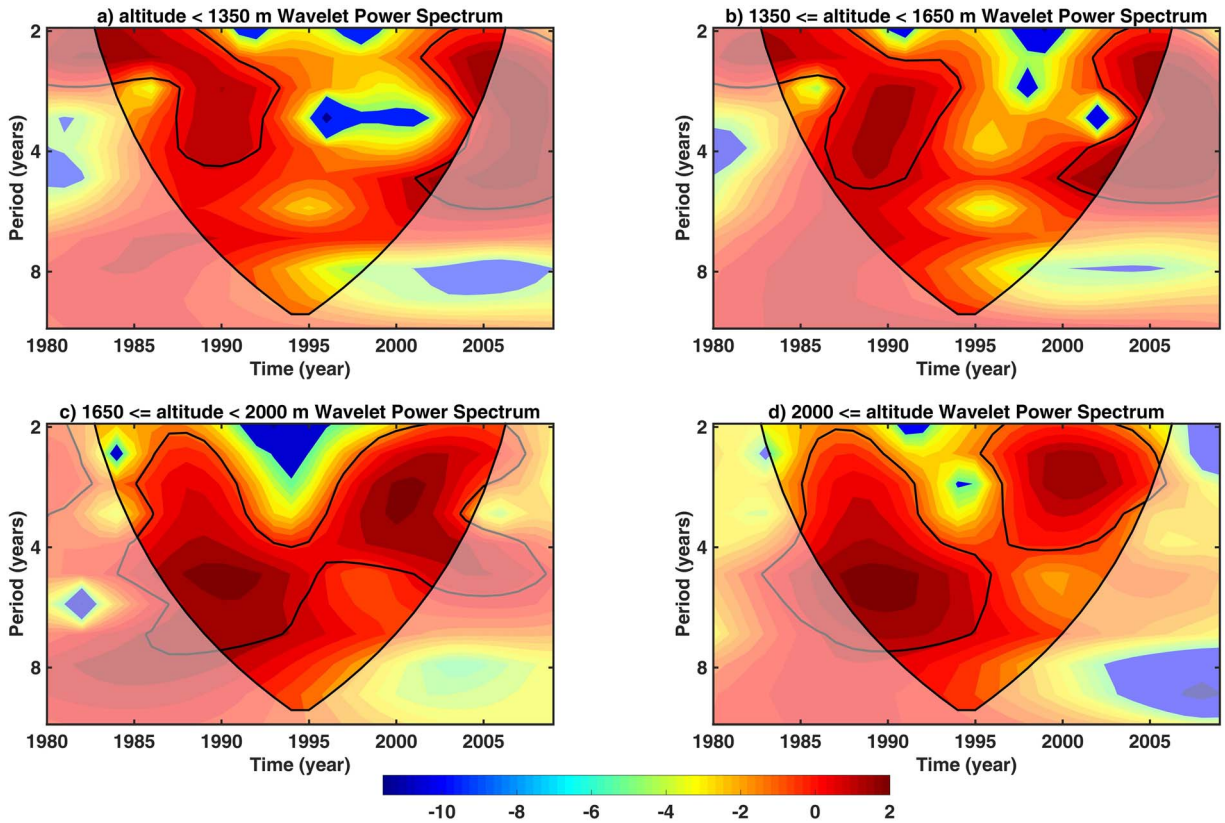
Fig. 6. Wavelet spectrum of the average snow cover for different altitude classes.
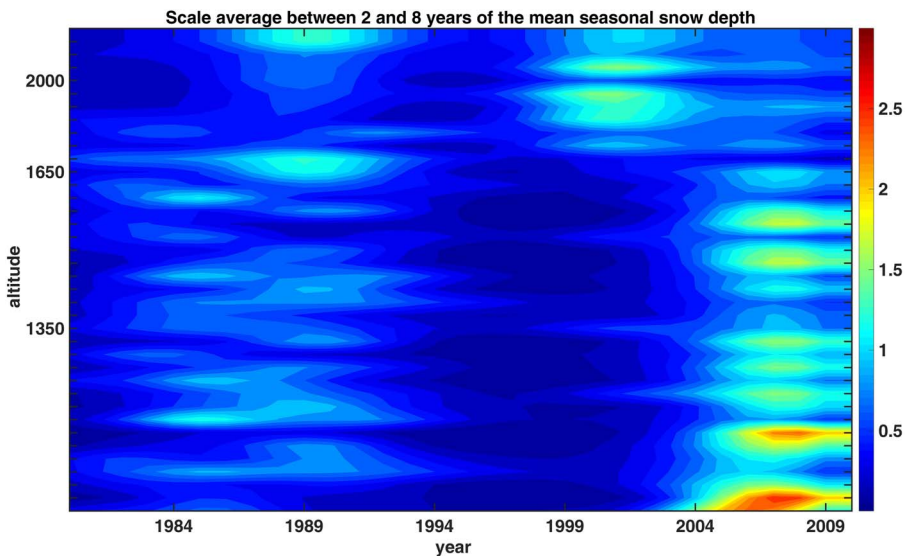

Fig. 7. Scale average between 2 and 8 year periods of mean seasonal snow depth.

the 1990s. Also high elevation sites show a peak after the 1990s, but it starts around 2000, hence anticipating the one observed for stations below 1650 m a.s.l., which starts after 2003. Overall, variations in the variance of the signal are more moderate at higher elevations. A similar pattern can be identified for snow cover duration, as shown in Fig. 8. The main difference can be observed for high elevation sites, where the reduction of the variance occurred in the 1990s is less pronounced than in the analysis of the mean seasonal snow depth signal.

Coherence between mean seasonal snow depth at different elevations and the NAOI and the MOI is shown in Figs. 9 and 10,
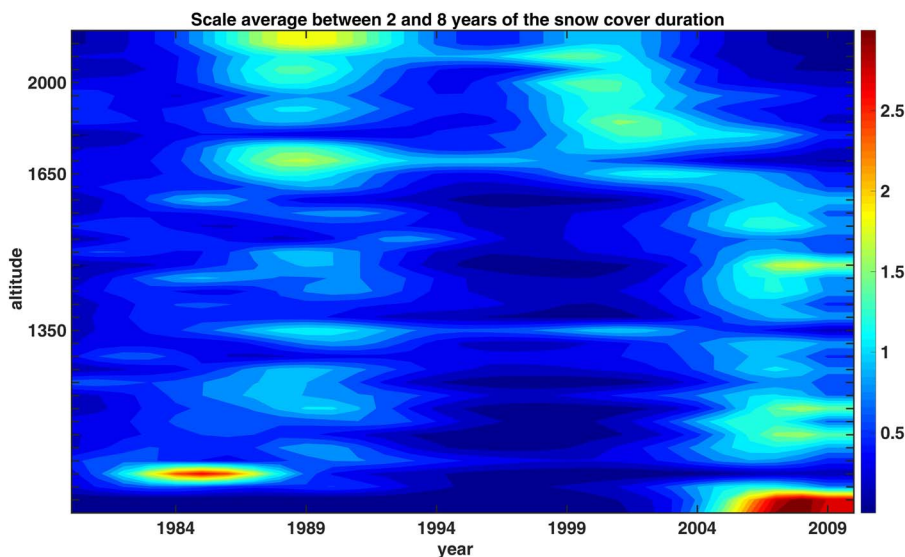
Fig. 8. Scale average between 2 and 8 year periods of snow cover duration.

respectively. The colors indicate the coherence of the two signals for different time and periods (red indicates high coherence, close to 1, blue indicates low coherence, close to 0) and the direction of the arrows indicates the relative phase between the signals (Grinsted et al., 2004). For example, if the signals are in phase the arrows point to the right and if the signals are in anti-phase the arrows point to the left.

We observe that the two lowest classes (below 1650 m a.s.l.) have a similar behavior, which is different than the behavior of the two highest classes (above 1650 m a.s.l.). The coherence with the NAOI is strong, i.e. close to 1, for all elevation classes up to the 2 year period, although for stations below 1650 m a.s.l. the coherence between the NAOI and the mean seasonal snow depth vanishes at the end of the 1990s (see Fig. 9a and b). Fig. 9a and b shows a strong coherence at 4 year period starting from 2000 for stations below 1350 m a.s.l. and starting from the 1990s for stations between 1350 m and 1650 m a.s.l. This signal is much weaker for the two elevation classes of stations above 1650 m a.s.l. (it is not even present in the end of the 1990s and the beginning of 2000s) and it is spread between the 4 and 6 year periods (Fig. 9c and d).

The coherence between the mean seasonal snow depth and the MOI shows that the highest stations (above 1650 m a.s.l.) display a strong coherence for the entire period from 1980 to 2009 between 6 and 4 year periods (see Fig. 10c and d). The coherence weakens at the end of the 1990s, but it is persistent. Also the lowest stations (Fig. 10a and b) show a strong coherence before the 1990s at a 6 year period. As a remarkable difference to the highest stations, the coherence decreases at the end of the 1980s and increases again after 2000 displaying a shift towards a shorter period of 4 years. In Figs. S2 and S3 we show that SCD displays a similar behavior.

4. Discussion

4.1. Difference in the behavior between sites located below and above 1650 m a.s.l.

The 1990s have been particularly dry and warm in the Alps (Brunetti et al., 2006b, 2009), which reduced both mean seasonal snow depth and snow cover duration as well as the probability of occurrence of precipitation in form of snow (Bartolini et al., 2009, 2010). Our results show that the effect of increasing temperature on the snow pack are strongly dependent on elevation. The variability in both mean seasonal snow depth and snow cover duration indicates that areas below 1650 m a.s.l. in the Adige catchment are significantly more affected by these changes than sites above 1650 m a.s.l. The variability in HS and SCD observed in high elevation sites therefore may be explained considering other important factors, such as wind and occurrence of solid pre-cipitations (Bartolini et al., 2009). Due to the lack of an adequate number of meteorological stations with high quality data at high elevations, it was not possible to carry on this analysis.

The elevation of 1650 m a.s.l. is within the sensitive elevation range observed by Beniston (1997) and by Laternser and Schneebeli (2003) for the Swiss Alps. However, notice that seasons with a particularly small mean seasonal snow depth and short snow cover duration, such as 1989 and 2001 (see Fig. 2), require actions (e.g. the use of artificial snow) to avoid negative impacts on touristic activities even at elevations higher than 1650 m a.s.l. This may have important implications for management of water resources and energy, since production of artificial snow consumes a significant amount of both resources (Rixen et al., 2011).

The wavelet analysis showed a difference between low and high elevation sites also in periods longer than one year (Figs. 5 and 6). In particular, it is possible to notice a difference in the wavelet power spectrum for low and high elevation sites both in case of mean seasonal snow depth and snow cover duration, especially at the 2 and 4 year periods. Moreover, the coherence to climatic indices also displays a remarkable dependence from the elevation (Figs. 9 and 10). In particular, we have observed a difference in the
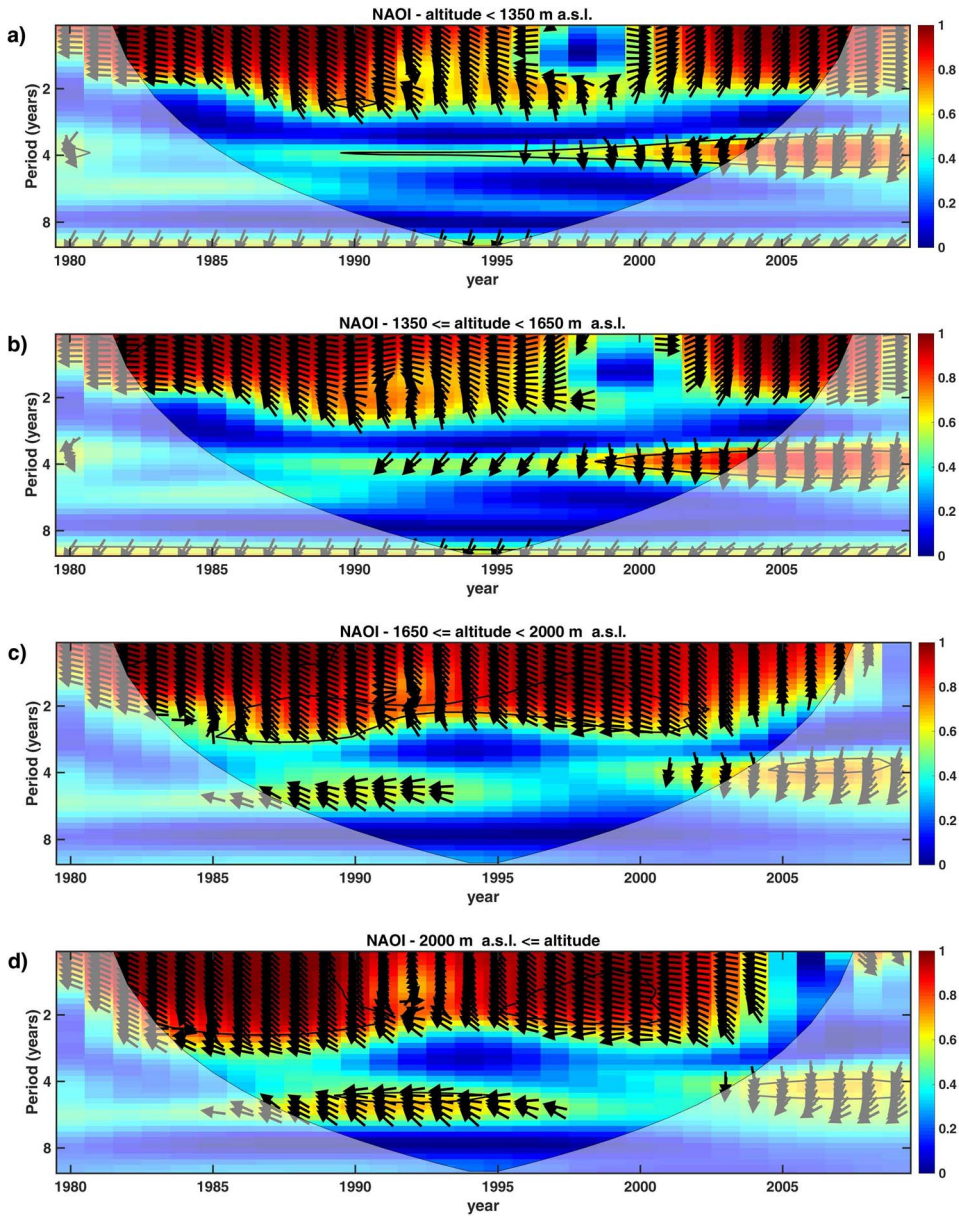
Fig. 9. Wavelet coherence of mean seasonal snow depth with the NAOI. The direction of the arrows indicates the relative phase between the signals.

coherence between HS and MOI and between HS and NAOI at 4 and 6 year periods at different elevations. This implies that sites located at different elevations show a different response to climatic changes and teleconnection patterns.

4.2. Difference in snow cover duration and mean seasonal snow depth analysis

Snow cover duration is influenced by the frequency of snowfall but also by the mean seasonal temperature. For example, an early snowfall event followed by a long cold and dry period may lead to an extended snow cover duration even if the number of snowfalls in the remaining part of the season may be very low or the events may be of low intensity. On the contrary, frequent snowfall events occurring in late winter or early spring, when temperature can rapidly increase during the day, do not lead necessarily to an extended snow cover duration. On the contrary, the mean seasonal snow depth is an indicator of the average amount of snow present at a site. This explains the differences observed in the behavior of these two variables, in particular at high elevations (see e.g. Fig. 2). The lower temperatures at high elevations (Fig. S1) allows to maintain the snow pack for longer periods than at lower elevations, even for seasons where the mean seasonal snow depth displays highly negative fluctuations (Fig. 2c). These negative fluctuations have to be explained by factors other than temperature (e.g. a decrease in solid precipitation). Moreover, different from mean seasonal snow depth, snow cover duration has a maximum value by definition of 182 days. Once this maximum value is reached, snow cover
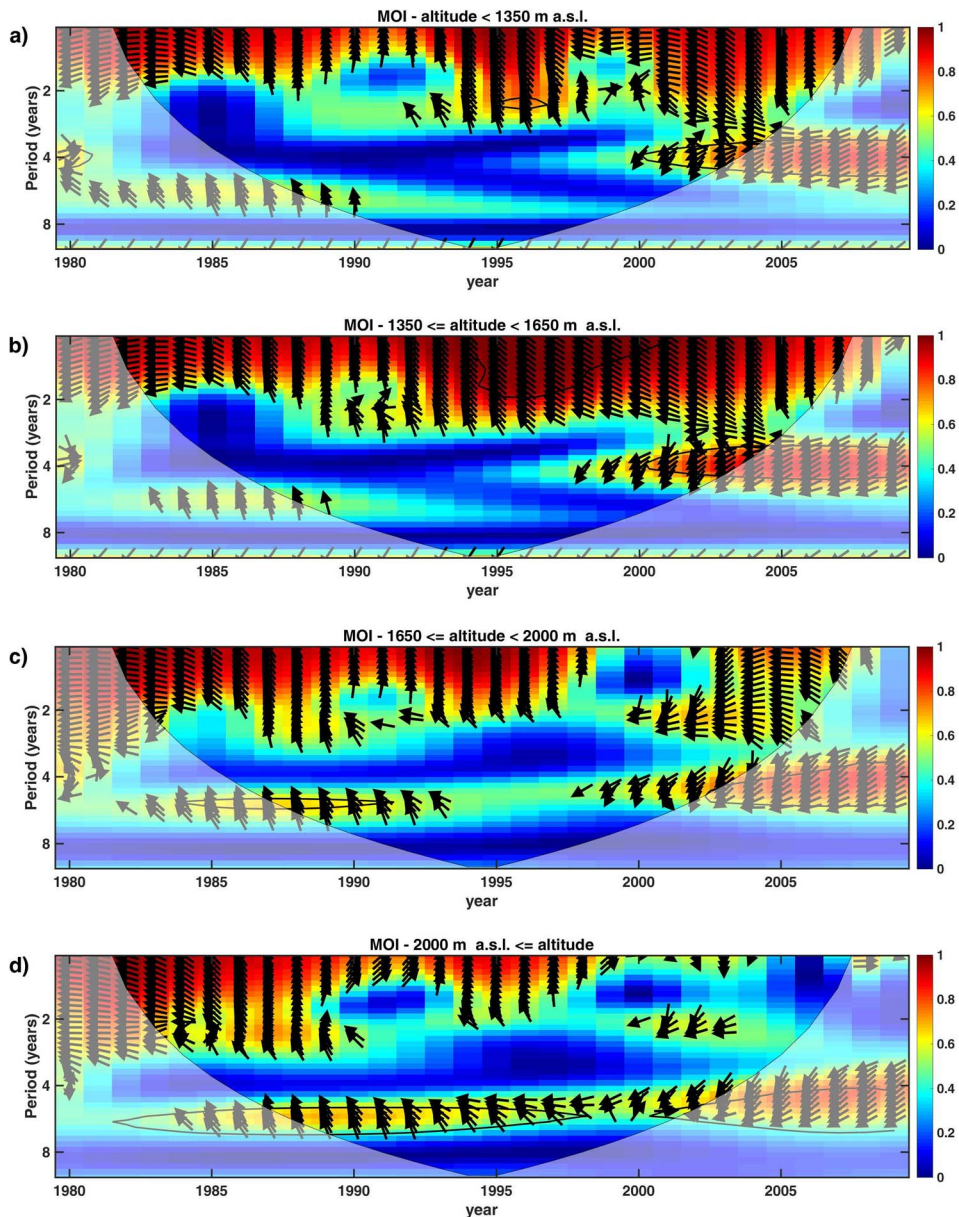
Fig. 10. Wavelet coherence of mean seasonal snow depth with the MOI. The direction of the arrows indicates the relative phase between the signals.

duration cannot differentiate between years where highly positive fluctuations in mean seasonal snow depth are observed. Snow cover duration and snow depth therefore bring complementary information about snow dynamics. Both are useful indicators depending on the research questions to be addressed (e.g. Keller et al., 2005; Groffman et al., 2001; Steiger and Stötter, 2013; Beniston, 2012a).

4.3. Difference in snow variability from 1988

The results of our analysis are in agreement with the findings of Marty (2008). Indeed, a period of about 18 years of negative fluctuations in HS and SCD begins in 1988 in the Adige river basin (Fig. 2). This change in time series has been confirmed by the analysis of Marcolini et al. (2017) who found around 1988 a natural change in most time series of mean seasonal snow depth available in the Trentino province, which covers most of the southern part of the Adige catchment. Beniston et al. (2003) observed a similar behavior for HS in the Swiss Alps, while Valt and Cianfarra (2010) reached similar conclusions observing time series of stations East and West of the observed area. Our study shows that, despite the heterogeneity in the local climatic conditions in the Alps (e.g. Brunetti et al., 2006b), the mean seasonal snow depth is low after 1988 also for the Southern ridge of the Alps. The

variability in mean seasonal snow depth is mainly controlled by the interplay between the observed increasing temperatures and the atmospheric pressure in the Alps (Brunetti et al., 2009). The correlation between temperature anomalies and mean seasonal snow depth shown in Fig. 4 therefore supports the hypothesis that temperature is one of the main driving factors of the observed change in snow dynamics in the Alpine Region in particular at low elevation sites (Scherrer et al., 2004).

The observed snow cover duration in our study area is consistent with the observations performed both in other sites in Northern Italy (Valt and Cianfarra, 2010) and Switzerland (Laternser and Schneebeli, 2003; Marty, 2008). Also for this variable, the short term (seasonal) variability strongly depends on local site specific conditions (Marty, 2008). However, our study shows that long term variability is controlled by global climate forcing factors, common for different regions of the Alps (e.g. Beniston, 1997; Marty, 2008; Valt and Cianfarra, 2010).

In order to evaluate the correlation of the variability of mean seasonal snow depth to global climate forcing factors, we analyzed its coherence with two oscillation indices: the NAOI and the MOI (Figs. 9 and 10). We cannot observe a clear relationship between the variations of the coherence between the NAOI and mean seasonal snow depth in the Adige River Basin and the changes observed in the behavior of snow in the last decades at periods longer than 2 years. In particular, the patterns observed for different elevations are different and do not reproduce the behavior observed for the anomalies in mean seasonal snow depth (decrease in the 1990s and increase, although not statistically significant, since 2000). The behavior shown in Fig. 10, instead, shows a change in the coherence between mean seasonal snow depth and the MOI at periods of 4–6 years during the 1990s. This is in accordance with the patterns observed for both mean seasonal snow depth and snow cover duration. The modification of this coherence during the 1990s, in correspondence with the period of the decrease of mean seasonal snow depth and snow cover duration, suggests a correlation between the MOI and the reduction of snow in the analyzed time period. The phase shift indicates that when the MOI and the snow signal display a strong coherence, they are almost anticorrelated. Moreover, the intensity of this change is stronger for stations below 1650 m a.s.l., confirming the higher sensitivity of these sites towards climate changes. In fact, for those stations we observe a break of the coherence during the 1990s, while for stations above 1650 m a.s.l. the coherence only decreases in this decade. In particular, the elevation dependence of the coherence between the mean seasonal snow depth and the MOI is highlighted in Fig. 10, where for the scales of 2 and 4 years the coherence for the stations below 1650 m a.s.l. reaches values close to 0.3 or even smaller than 0.2, while for stations above 1650 m a.s.l. the coherence does not reach values smaller than 0.5.

5. Conclusions

In this article, we analyzed the snow depth time series in the Adige catchment (North-East Italy) for the period from 1980 to 2009. Our results show that the Adige catchment experienced a significant reduction of both snow cover duration and mean seasonal snow depth after 1988. This reduction is observed both at low and high elevation sites, although we can distinguish between the behavior of the time series located above and below 1650 m a.s.l. In particular, low elevation sites are more affected by climate variability and more sensitive to temperature increase than high elevation sites. The difference between high and low elevation sites is also confirmed by the wavelet analysis. We also observe that in the analyzed period the coherence between mean seasonal snow depth signal and the MOI is low when mean seasonal snow depth is low at the scale between 4 and 8 years.

Due to the relevance of snow dynamics for other environmental processes, this study can provide an important reference for the explanation of changes, such as temporal shifts in streamflow peak, touristic and chemical fluxes, observed in this region in the past decades.

Conflict of interest

All authors declare that they have no conflict of interest.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.ejrh.2017.08.007.

Appendix A. Wavelet analysis

The continuous wavelet transformation of a discretized signal x_n, $n = 1, ..., N$, sampled at the time interval δt, is defined as

follows (Grinsted et al., 2004):

$$W_n(s) = \sum_{n'=0}^{N-1} x_{n'} \psi^* \left[\frac{(n'-n)\delta t}{s} \right]$$

(A.1)

where $\psi(\eta)$ is a band-pass filter and the superscript * indicates the complex conjugate. Mathematically, the transform in Eq. (A.1) is the convolution of the signal x_n with the scaled version of the wavelet function ψ, which is obtained from a "mother wavelet" $\psi_o(\eta)$, normalized at each scale s to fulfill the following condition:

$$\sum_{k=0}^{N-1} |\psi(s\,\omega_k)|^2 = N$$

(A.2)

where $\omega_k = 2\pi k/(N\delta t)$ for $k \leq N/2$, and $\omega_k = -2\pi k/(N\delta t)$ for $k > N/2$ is the angular frequency and s is the wavelet scale (Torrence and Compo, 1998). In this study, the Morlet function was used as "mother wavelet" and it is defined as:

$$\psi_o(\eta) = \pi^{-1/4} e^{i\,\omega_o \eta} e^{-\eta^2/2}$$

(A.3)

where ω_o is the dimensionless frequency (Grinsted et al., 2004) and η is the dimensionless time. Following Torrence and Compo (1998) ω_o is set to 6. In this way the wavelet scale is almost identical to the corresponding Fourier period.

The wavelet transform of Eq. (A.1) is computed for a selection of scales:

$$s_j = s_0^{2j\delta_j}, \quad j = 1, 2, ..., J$$

(A.4)

where s_0 is the smallest scale considered in the analysis and $J = \delta_j^{-1} \log_2(N\delta_t/s_0)$ determines the largest scale. Moreover, δ_j is the inverse of the number of scales per each octave, i.e. the distance between one frequency and its double. In addition, s_0 should be chosen in order to make the equivalent Fourier period approximately 2 δt.

The wavelet power spectrum (WPS) is defined as the product of wavelet transform, $W_n(s)$, by its conjugate, $W^*_n(s)$.

The Wavelet coherence (WTC) between two time series X and Y with wavelet transforms $W_n^X(s)$ and $W_n^Y(s)$, respectively, is defined as (Torrence and Compo, 1998; Grinsted et al., 2004):

$$R_n^2(s) = \frac{|S[s^{-1} W_n^{XY}(s)]|^2}{S[s^{-1} |W_n^X(s)|^2] S[s^{-1} |W_n^Y(s)|^2]}$$

(A.5)

where $W_n^{XY}(s) = W_n^X(s) W_n^{Y*}(s)$, where the superscript * indicates the complex conjugate, S is a smoothing operator, given by

$$S(W) = S_{\text{scale}}\{S_{\text{time}}[W_n(s)]\}$$

(A.6)

where S_{scale} denotes smoothing with respect to the scales s and S_{time} smoothing in time. For the Morlet wavelet Torrence and Compo (1998) suggested the following smoothing operator:

$$S_{\text{time}}(W)|_s = (W_n(s)*c_1^{-t^2/2s^2})|_s,$$
$$S_{\text{time}}(W)|_s = (W_n(s)*c_2 \Pi(0.6\,s))|_n$$

(A.7)

where c_1 and c_2 are normalization constants and Π is the step function. The factor 0.6 is the empirically determined scale decorrelation length for the Morlet wavelet (see Torrence and Compo, 1998).

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