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REGIONAL ESTIMATION OF SNOW WATER EQUIVALENT IN THE ITALIAN ALPS USING KRIGING

ABSTRACT: BOCCHIOLA D., *Regional estimation of Snow Water Equivalent in the Italian Alps using Kriging.* (IT ISSN 0391-9838, 2010).

Spatial estimation of Snow Water Equivalent at snowmelt is tackled using kriging from a sparse network of 34 snow stakes within the central Italian alps and pre-alps, for the period 1985-2005. First, sparse snow density pits first are used to find a relation between snow density, geomorphologic features and the season. Then, density estimates are used to evaluate SWE at snow stakes. To highlight regional features in SWE distribution, a Principal Components Analysis (PCA) is carried out, resulting into partition of the investigated area in two different homogeneous regions featuring different SWE statistics, consistent with previous findings. Then, the covariance of the SWE field within the two regions is studied, necessary for kriging, and its regularization provided based upon geomorphic attributes. Then, a Kriging procedure based upon the so obtained covariance fields is developed and cross-validated. Kriged SWE maps are then illustrated for two sample years, to demonstrate use of the method. The procedure provides well estimated, least variance SWE values and it is relatively simple and fast because it uses only information upon altitude at the investigated sites. Therefore, it can be used for spatial estimation of SWE within the investigated region for hydrological conjectures.

KEY WORDS*:* Snow Water Equivalent, Kriging, Alps.

RIASSUNTO: BOCCHIOLA D., *Stima regionale dell'Equivalente Idrico Nivale nelle Alpi Italiane tramite procedura di Kriging.* (IT ISSN 0391-9838, 2010).

Si affronta la stima dell' Equivalente Idrico Nivale (EIN) al disgelo tramite procedura di kriging a partire da una rete di 34 stazioni di misura del manto nevoso nelle Alpi e pre Alpi centrali Italiane, per il periodo 1985-2005. Per prima cosa, si utilizzano rilievi nivologici sparsi per determinare una relazione tra la densità della neve, le caratteristiche geomorfologiche dei siti ed il periodo. Quindi, le stime di densità vengono utilizzate per calcolare l'EIN nelle stazioni. Per evidenziare la presenza di effetti regionali nella distribuzione dell'EIN, si conduce un'analisi alle componenti principali (PCA), che fornisce la ripartizione della regione investigata in due regioni omogenee contraddistinte da differenti proprietà statistiche dell'EIN, in accordo con studi precedenti. Quindi, il campo di covarianza dell'EIN nella regione viene studiato, come è necessario per la procedura di kriging, e regolarizzato tramite una funzione degli attributi geomorfologici. Infine si conduce la procedura di kriging sulla base delle covarianze così ottenute e si sviluppa una procedura di validazione incrociata delle stime di EIN ottenute. Si producono poi le mappe di EIN tramite kriging per due anni campione, per illustrare l'applicazione del metodo. La procedura sviluppata fornisce valori di EIN stimati in maniera ottimale e di minima varianza ed è relativamente semplice e veloce, poiché utilizza in pratica la sola informazione relativa all'altezza dei siti studiati. Il metodo può quindi essere utilizzato per la stima spaziale dell'EIN nell'area di studio allo scopo di condurre studi idrologici.

TERMINI CHIAVE: Equivalente Idrico Nivale, Kriging, Alpi.

INTRODUCTION

The recognized evidence of global warming demands for assessment of water resource from the cryosphere in temperate regions, and more challenging foresight of its destiny, including potential extinction of permanently glacierized areas, and the tremendous impact upon the Alpine environment of modified seasonal snow cover (Barnett & *alii*, 2005). In Italian mountain ranges importance of water from cryosphere has emerged clearly during latest dry summers, most notably in year 2003, when it saved most of the tributaries of the Po river from the severest droughts, and evidence is raising of ongoing variability of European Alpine water resources due to transient climate change (Rohrer & *alii*, 1994; Beniston, 1997; Laternser & Schneebeli, 2003; Bianchi Janetti & *alii*, 2008; Gorni & *alii*, 2008; Maragno & *alii*, 2009. While snow cover extent, duration and dynamics influence vegetal and animal biota

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in Alpine areas (*e.g.* Erschbamer, 1989; Gottfried & *alii*, 1999; Theurillat & Guisan, 2001; Fiorese & *alii*, 2006; Keller & *alii*, 2005), freshwater availability from cryosphere during spring and summer regulates hydrological cycle of Alpine basins, and influences Alpine ecosystems development (*e.g.* Coughlan & Running, 1997, McGlynn & *alii*, 1999, Maiolini & Lencioni, 2001; Beniston & *alii*, 2003). Water budget of the cryosphere is driven, on one side, by snow cover forming during winter, and by its redistribution by gravity and wind (*e.g.* Jansson, 1999; Kuhn, 2003; Wagnon & *alii*, 2007), and on the other side, by energy budget of snow and ice, leading to evaporation and ablation (*e.g.* Horne & Kavvas, 1997; Singh & *alii*, 2000; Lehning & *alii*, 2002; Hock, 2003, 2005). Snow cover delays ice melting and crevassing upon glaciers (*e.g.* Diolaiuti & *alii*, 2006) and snow surviving summer thaw at high altitudes eventually feeds ice cover in glacierized areas (*e.g.* Hock, 2005). As a result of the several processes governing snow cover dynamics, snow cover patterns show a tremendous variability in space and time (Tani, 1996; Blöschl, 1999, Yuang & Woo, 1999; Lehning & *alii*, 2008; Manes & *alii*, 2008) and their proper assessment at both macro and micro-scale is of utmost importance (*e.g.* Anderton & *alii*, 2002). As such, any modelling approach to water and ice budget within alpine areas requires accurate assessment of Snow Water Equivalent (SWE) at thaw start (*e.g.* Coughlan & Running, 1997, Carrol & Cressie 1997, Elder & *alii*, 1998), conventionally set at April 1st (thus named insofar SWE_1, see *e.g.* Ranzi & *alii*, 1999, Bohr & Aguado, 2001), although this date can slightly change from site to site. Albeit so, SWE_1 estimation for water budget purposes is normally carried out using very simple averaging methods (*e.g.* Ranzi & *alii*, 1991), the main drawback of which is the (lack of) assessment of the accuracy of estimation, which remains unknown. When dealing with SWE estimation, a crucial issue is snow density appraisal. Density is some way more conservative as compared to depth, particularly after snow settlement and ripening (*e.g.* Elder & *alii*, 1998) and requires a minor amount of measurements, while its tracking is time consuming and unwieldy and seldom comprehensive density data are available. As such, empirical rules for snow density estimation are usually introduced (*e.g.* Elder & *alii*, 1991; Ranzi & *alii*, 1999). When considering large areas, regional issues may be considered (*e.g.* Elder & *alii*, 1991, Bohr & Aguado, 2001). These allows gathering of information from all the available data set over a wide area, by correctly identifying zone featuring similar (homogeneous) properties (*e.g.* Bocchiola & *alii*, 2006; Bocchiola & Rosso, 2007a, b; Bocchiola & *alii*, 2008). An optimal estimation method for spatially distributed variables is kriging, which is BLUE (Best Linear Unbiased Estimator), and provides accuracy of the estimates (*e.g.* Davis, 1973; Cressie, 1993) and has been already tentatively applied to SWE estimation (Carrol, 1995; Carrol & Cressie, 1997). Here, I deal with a regional approach to kriging of SWE for a wide snow-covered area in Lombardia region, Northern Italy, not attempted insofar in my knowledge, unless by Martinelli & *alii*, 2004. I here modify and improve the results therein by using more gage stations and considerable more years. This should allow to get better estimates of the regional SWE.

I rely here upon measurements of snowpack depth from 34 automatic stations from trained personnel of the regional environmental protection agency, ARPA Lombardia, for the period 1985-2005. The area in study includes a number of major river basins, featuring relevant snowmelt component in spring-summer water resources recruitment, necessary for anthropic uses as well as for riverine ecosystems maintenance, and thus strongly depending upon water budget from the cryosphere. Because no snowpack density measurements are available at the gauging stations, I use detailed snow properties measurement from a sparse data base made available by AINEVA association to find a link between snow density and the main geo-morphologic features of the sites and the season.

I then carry out a Principal Component Analysis (PCA, *e.g.* Wootling & *alii*, 2000, Bohr & Aguado, 2001; Rohrer, 1994) to highlight regional features of SWE_1, based upon former works labelling this approach as warranted (*e.g.* Bocchiola & *alii*, 2006; Bocchiola & Rosso, 2007b). Two homogeneous regions are highlighted, where relevant difference in the SWE_1 statistical properties are observed. To carry out spatial interpolation of point measured SWE_1 by Kriging technique, I then investigate the correlation structure of the spatial SWE_1 field from the available observations. To cope with covariance estimation in ungauged sites, I model the observed correlation field as isotropic, with its values depending upon some geo-morphologic features of the observation sites (*e.g.* Lapen & *alii*, 1996, Carrol & Cressie, 1997). Time average and standard deviation of SWE_1 at ungauged sites are assessed by using correlation analysis with geomorphic and climatic attributes. To test accuracy of the kriging method, I use cross validation to back calculate SWE_1 at each station from measurements at the other ones and evaluate some statistical performance indicators. Eventually, I illustrate application of the method to production of kriged SWE_1 maps for two sample years.

CASE STUDY AREA AND AVAILABLE DATA BASE

The study area (fig. 1) covers the alpine part of Lombardia region, situated in Central-Northern Italian Alps. This includes four major river basin, namely Adda, Brembo, Serio and Oglio river basins. This amounts to an area of about 6250 km^2 above 800 m asl, covered by a network of 34 snow gauges, for a density of snow gauges of one very 184 km^2. A someway higher density of snow gauges is spotted in the central part of the area, *i.e.* upon the Orobie pre-alps and in the North Eastern part, Alta Valtellina. The measuring stations (and data base) are properties of ARPA Lombardia, Milano (see tab. 1 for main stations' characteristics). The automatic stations provide daily snow depth measurements over a series of years, from 1985 to 2005 (but data of year 2001 are completely missing), together with a number of meteo data. Eventually, I could use a number of 488 snow depth measurements.

FIG. 1 - Digital Elevation Model (DEM) of the area investigated in the present study. The two regions highlighted via PCA are indicated.

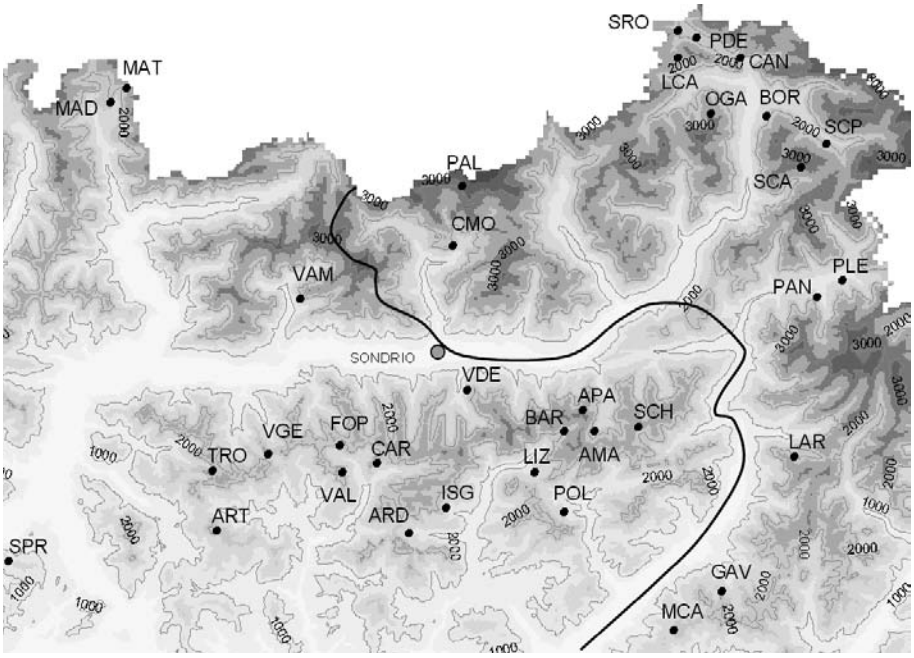

TABLE 1 - Available data base. *Ea* is East coordinate, *No* North coordinate, *A* is altitude and n_s the number of available years. Italic indicates stations with less than 15 surveys. The sample mean and variance of estimated SWE_1 are reported

Site	Station	*Ea* [m]	*No* [m]	*A* [m asl]	n_s [.]	μ_{SWE1} [cm]	σ_{SWE1} [cm]
Aprica Magnolta	AMA	1588734	5109119	1950	19	33.00	10.84
Aprica palabione	APA	1590431	5108985	1920	*8*	49.33	20.98
Ardesio	ARD	1566246	5089668	1480	*4*	18.53	13.99
Artavaggio	ART	1541010	5086858	1635	19	17.59	21.01
Barbellino	BAR	1580562	5101767	1850	19	33.55	25.07
Bormio 2000	BOR	1606800	5143975	2010	20	19.40	9.80
Cancano	CAN	1601039	5152028	1950	20	22.55	12.46
Carona	CAR	1561214	5094295	1840	19	28.94	23.02
Campo Moro	CMO	1571450	5128154	1970	19	18.40	13.27
Foppolo	FOP	1558893	5099179	1680	*13*	23.90	16.38
Gaver Bagolino	GAV	1613068	5085470	1450	*14*	14.10	11.08
Spiazzi di Gromo	ISG	1575533	5089868	1638	*4*	30.89	5.56
Lago d'Arno	LAR	1610673	5100416	1830	*11*	27.79	18.47
Livigno Carosello	LCA	1583519	5152057	2675	18	80.62	28.61
Lizzola	LIZ	1577756	5096726	1490	17	22.56	15.26
Madesimo	MAD	1527201	5142067	1530	20	22.56	15.39
Madesimo Mater	MAT	1528345	5141572	1860	20	27.45	17.03
Montecampione	MCA	1596390	5076789	1765	19	17.23	12.60
Oga S.Colombano	OGA	1600337	5145317	2295	*4*	41.02	13.85
Palù	PAL	1568142	5127000	2155	*9*	34.85	19.23
Pantano	PAN	1613698	5114798	2105	19	37.07	22.47
Passo d'Eira Mottolino	PDE	1589472	5154542	2395	*4*	28.70	18.43
Ponte di Legno	PLE	1616110	5128155	1600	16	4.54	7.05
Colere Monte Polzone	POL	1581663	5092306	1570	20	20.53	20.29
Plaghera	SCA	1614256	5139449	2240	20	23.79	11.35
Schilpario	SCH	1589995	5095552	1150	*4*	7.00	6.46
S. Caterina Valfurva	SCP	1615320	5140769	1735	*4*	6.29	9.28
San Primo	SPR	1517717	5085709	1070	19	4.07	5.40
Livigno S. Rocco	SRO	1586394	5152844	1875	20	12.21	10.80
Trona	TRO	1541685	5097781	1810	19	31.54	19.32
Valleve S. Simone	VAL	1554917	5099064	1660	20	19.88	17.19
Valmasino	VAM	1549358	5118642	855	16	1.36	3.48
Vedello	VDE	1569577	5106220	1060	*4*	23.72	10.84
Valgerola	VGE	1544339	5097306	1840	20	32.88	16.99

Snow station altitude ranges from 855 m asl at VAM (Val Masino), to 2675 m asl at LCA (Livigno Carosello). Snow density is not routinely measured at the automatic stations. Therefore, SWE measurements are not directly available, but can only be inferred coupling depth measurements with density estimation based upon some *ad hoc* developed approach. To this purpose, I gathered data from 196 snow surveys carried out during 1996-2002 by Italian snow and avalanche association (AINEVA) personnel in sparse sites over the investigated area (not shown, reported in Martinelli & *alii*, 2004), providing, among other variables, vertical profiles of snow density and temperature.

UNKNOWN SNOW DENSITY ESTIMATION

I here rely upon a formula preliminarily developed by Martinelli & *alii* (2004) to estimate (depth) averaged snow density at a given site, depending upon some morphologic and climatic features and the period of the year (*e.g.* Elder & *alii*, 1991). In short, using forward stepwise regression Analysis (FSR) (*e.g.* Kottegoda and Rosso, 1997, p. 340, carried out using EXCELS® XLStat® package, also used for all regression analysis here) we found out four most significant variables (at 5% significance level, Student's T test), namely the number of days past from the 1st of September *D*, the altitude *A* in meters, the site aspect (–180° to 180°) *AS*, the local slope (0° to 90°) *I* and the average snow temperature (Celsius degrees, negative values) *T*. Snow density [kg m^{-3}] can be thus estimated as

$$\hat{\rho}_s = 134 + 0.54 \cdot D + 0.06 \cdot A + 0.10 \cdot AS - 1.45 \cdot I + 6.09\ T\ . \quad (1)$$

TABLE 2 - Regression analysis (lines 1-2, see Section 2.2, lines 3-6, see section 2.4, line 7 see section 2.7). Symbols explained in text. N is sample numerosity (X-Y couples). *p-val*< 0.05 indicates significance of the regression

Y/X	N [.]	*Interc.* [Y]	A [m asl]	D [.]	AS [°]	I [°]	T [°C]	DA [m]	*p-val* [.]	R^2 [.]
$?_s$ [kgm^{-3}]	196	134	0.06	0.54	0.10	-1.45	6.09	–	<0.0001	0.56
$?_s$ [kgm^{-3}]	196	129	0.04	0.64	0.11	-1.47	–	–	<0.0001	0.51
μ_{SWE1} S-W [cm]	19	-25.59	0.031	–	–	–	–	–	<0.0001	0.93
μ_{SWE1} N-E [cm]	15	-41.27	0.031	–	–	–	–	–	<0.0001	0.61
$?_{SWE1}$ S-W [cm]	19	-8.15	0.014	–	–	–	–	–	<0.0001	0.78
$?_{SWE1}$ N-E [cm]	15	-17.95	0.016	–	–	–	–	–	<0.0001	0.70
Ln($?_{i,j}$) [.]	210	-0.28	–	–	–	–	–	-0.0009	<0.0001	0.29

The formula provides a regression coefficient $R^2 = 0.56$ (see tab. 2, scatter plot agreement not shown here, reported in Martinelli & *alii*, 2004). The choice of the variable D arises because no snow depth was observed in any gauged site before September 1st. The geo-morphologic variables used here have been calculated starting from the available digital elevation model, featuring a resolution (East, North) of 230 by 220 meters. Other variables have been included into the FSR procedure, *i.e.* snow height, vegetation cover and land use, but no improvement of the explained variance was obtained. Also, in view of the different measuring sites, it was not possible to consider local topography in the surveyed point, including holes, benches, ramps, etc. (as in *e.g.* Elder & *alii*, 1991). No direct information could be obtained about the incoming net radiation, possibly related to snow density (see *e.g.* Elder & *alii*, 1991). Besides, D is somewhat related to net radiation and could account for it indirectly (see Elder *et al.*, 1991), but at the cost of a minor effort. The determination coefficient R^2 (and thus the predictive power) of Eq. (1) is somewhat low. However, in view of the complexity of the investigated phenomenon and the size of the considered area, the results are considered reasonably good (see, for a comparison, Elder & *alii*, 1991, Elder & *alii*, 1998). We tentatively tried to highlight any regional features in the snow density dependence upon the observed variables, but we could not find any. One notices that the date of the survey is the most important feature influencing snow density. This makes sense, because if snow melt does not occur, density variation is mainly due to settling phenomena in the snowpack, which under its proper weight decreases its depth, thus increasing its density in time (for a settlement model based on the number of days from the snow fall, see *e.g.* Martinec & Rango, 1991). When considering April 1st, one has $D = 212$ in Eq. (1). Here SWE_1 is estimated by the observed snow depth h_{s1} times the estimated snow density, $SWE_1=?_{s1}h_{s1}$. Because the ARPA automatic snow stations do not provide snow temperature, this is not used here. Although this SWE_1 value is affected by some amount of error, due to uncertainty in snow density estimation, I here take SWE_1 as a correct one for the purpose of the further developments, in the hypothesis that noise given by snow density estimation is negligible against the noise in SWE_1 estimation at ungauged sites. In Section 3.4, I discuss the influence of uncertainty of $?_{s1}$ estimation upon kriging of SWE_1.

REGIONAL FEATURES IN SWE DISTRIBUTION

I evaluate here relevant regional features in the properties of SWE_1 distribution, related to the geo-morphologic and/or climatic properties of the area (*e.g.* Wootling & *alii*, 2000). Statistical grouping techniques can be adopted for this purpose, including Principal Components Analysis (PCA), Factor Analysis (FA) and Cluster Analysis (CA) (*e.g.* Rohrer & *alii*, 1994; Baeriswyl and & Rebetez, 1997; Bocchiola & Rosso, 2007b; Bocchiola & *alii*, 2008, also currently used for regionalization of snow properties. I carried out several attempts, using PCA, CA, FA (also with Varimax rotation) and also CA of the resulting loadings (*e.g.* Baeriswyl & Rebetez, 1997; Bohr & Aguado, 2001). However, the most reasonable results were obtained by PCA. Here, I carried out direct PCA of SWE_1, conversely to what reported in Martinelli & *alii* (2004), where the cumulated yearly amount of water in fresh fallen snow, SWE_Y was used. In facts, the correspondence between SWE_Y and SWE_1 (investigated *e.g.* by Bohr & Aguado, 2001) was observed to be statistically valid here above an altitude of about 1700 m asl, whereas for lower altitudes such hypothesis breaks down (not shown here, reported in Martinelli & Modena, 2003). Because several stations here are placed below that altitude, I directly investigate SWE_1, which is of interest here as an initial condition for thaw season. For robustness, I used for PCA only stations featuring at least 15 years of data, resulting into 21 stations (see tab. 1). tab. 3 reports the first three components from PCA of SWE_1 (explaining 0.80 of the sample variability). The first component displays clearly the great-

TABLE 3 - PCA. Loadings on the First 3 Principal Components (Cum. Variability = 0.78). Greatest loadings in bold. Italic second greatest loading

Variability/Loading	PC1	PC2	PC3
Variability [.]	0.65	0.08	0.08
Cum. Variability [.]	0.65	0.72	0.80
AMA	**0.72**	0.12	*0.48*
ART	**0.86**	0.12	*-0.28*
BAR	**0.94**	0.02	*-0.23*
BOR	**0.80**	*-0.39*	0.08
CAN	**0.87**	*-0.44*	0.03
CAR	**0.94**	-0.03	*-0.18*
CMO	**0.95**	*-0.21*	-0.05
LIZ	**0.77**	*0.48*	-0.13
MAD	**0.78**	-0.03	*-0.15*
MAT	**0.91**	-0.13	*-0.17*
MCA	**0.77**	0.17	*0.55*
PAN	**0.53**	*0.45*	-0.24
PLE	**0.61**	-0.32	*0.36*
POL	**0.95**	*0.14*	-0.08
SCA	**0.71**	*-0.37*	-0.23
SPR	**0.57**	0.31	*0.62*
SRO	**0.87**	*-0.22*	0.05
TRO	**0.96**	0.07	*-0.10*
VAL	**0.97**	0.07	*-0.11*
VAM	**0.35**	-0.18	*0.39*
VGE	**0.72**	*0.53*	0.01

est loadings from all the stations, thus suggesting it represents general climate signature within the area. However, investigation of the loadings of SWE_1 upon the second and third components clearly indicates a consistent spatial pattern. Also, a two-factor analysis FA carried out upon SWE_1 provided a practically equivalent partitioning (not shown for shortness), so substantiating the results here. The obtained spatial grouping of the stations is reported in fig. 1. The two regions shown therein mirror topographic and climatic features of the observed area. Particularly, they are laid accordingly to a main axis oriented in the south-east to north-west direction, approximately following the main direction of the Adda river and Oglio river valleys. The two regions will be hereinafter referred to as South-West (Region 1, S-W) and North-East (Region 2, N-E). This partitioning is in practice coincident with that proposed by Martinelli & *alii* (2004), based upon SWE_Y as reported, as well as with that found by Bocchiola & Rosso (2007b), considering greatest yearly three-day snow fall H_{72} for avalanche hazard mapping. Comparison of these results seemingly indicate a consistent signature of climate and topography upon precipitation under snow form, and consistently upon available water at thaw. Some minor adjustment based upon geographical considerations were made, mainly along the boundary between regions, which may of course be considered approximate (compare *e.g.* with Bocchiola & Rosso, 2007b).

Notice that the Oglio river, in the East part of the region, is split in two parts. Although a single watershed is not expected to show regional differences inside its boundaries, here it appears that such differences may be justified in view of the orientation of the Oglio basin hillslopes, following a S-W to N-E direction. Again, a similar issue was observed in the former studies reported here. Positioning of snow measuring stations within the western slopes of the watershed may help to discriminate such issue in the future. The general findings related to partition here are also confirmed by the personnel of the AINEVA agency in Bormio, based on their personal qualitative experience, although no detailed study has been carried before (G. Peretti, personal communication, April 2003).

KRIGING OF OBSERVED SWE DATA

Kriging is usually carried out upon standardized data sets (*e.g.* Carrol & Cressie, 1997). For *SWE* data, this amounts to calculate

$$SWE^* = \frac{SWE - \mu_{SWE}}{\sigma_{SWE}}, \tag{2}$$

Where μ_{SWE} and σ_{SWE} are the mean and the standard deviation of the point site SWE. In ungauged sites, these can be estimated from local proxy data, including morphology or climate (see *e.g.* Carrol, 1995). The estimate of SWE_0^* at an ungauged site from the known values at n snow gauges is given by

$$S\hat{W}E_0^* = \sum_{i=1}^{n} SWE_i^* \lambda_i, \tag{3}$$

where the weights λ_i are assessed from the

$$\lambda = \Sigma^{-1} c, \tag{4}$$

where Σ is an n x n covariance matrix of the SWE_i at the gauged sites and c is a n x 1 column vector of the covariances among the SWE_i at the observed gauges and the ungauged one. For estimation of unknown SWE_0, the covariance matrices are obtained by regularization of observed covariances against morphologic characters, such as distance or altitude jump (normally using the hypothesis of «isotropic» covariance, *e.g.* Carrol & Cressie, 1997). The variance of estimation, necessary to quantify estimation accuracy for SWE^* is then given

$$\hat{\sigma}^{*2}_{\mathrm{S\hat{W}E\,0}} = 1 - c^T \Sigma^{-1} c. \tag{5}$$

The estimation variance for unstandardized SWE_0 is

$$\hat{\sigma}^{2}_{\mathrm{S\hat{W}E0}} = \sigma^{*2}_{\mathrm{S\hat{W}E0}} \sigma^{2}_{\mathrm{SWE\,0}}, \tag{6}$$

giving the expected accuracy of the SWE_0 estimate as the product of the estimation variance of SWE_0^* times SWE_0 variance, known or someway estimated.

ANALYSIS OF THE STATISTICAL PROPERTIES OF SWE$_1$ INCLUDING GEO-MORPHOLOGIC ATTRIBUTES

At unknown sites, information is required about mean and variance of SWE_1, necessary in Eq. (2) and Eq. (6) (*e.g.* Carrol, 1995). These can indeed be related to the main geo-morphologic and climatic properties of the sites. Here, I carried out a FSR regression analysis to identify the most significant (5% significance level) physical variables in the assessment of μ_{SWE1} and μ_{SWE1}. I considered *Ea* and *No* coordinates, altitude *A*, aspect *AS*, and slope *I*, and used weighted linear regression (upon estimation variance of μ_{SWE} and μ_{SWE}). However, only *A* was found to significantly impact μ_{SWE1}, as

$$\hat{\mu}^{\mathrm{S\text{-}W}}_{\mathrm{SWE1}} = -25.59 + 0.031 \cdot A, \tag{7}$$

with μ_{SWE1} in cm. The equation provides a determination coefficient $R^2 = 0.93$. For Region N-E, one has

$$\hat{\mu}^{\mathrm{N\text{-}E}}_{\mathrm{SWE1}} = -41.27 + 0.031 \cdot A, \tag{8}$$

with $R^2 = 0.61$ and again only *A* significant variable. Notice the same slope $s_A = 0.031$ for both regions, indicating a similar rate of increase. However, a considerable difference is seen in the intercept s_0, indicating on average 16 or so less cm of SWE_1 in region N-E than in region S-W (notice from tab. 2 that there is no overlapping between the two 5% confidence intervals for s_0 in the two regions). In fig. 2 it is clearly visible a shift between the two curves, indicating that in region N-E less snow is expected at April 1st than in region S-E. This effect gives reason of the regional grouping as given by PCA, which indicates clearly the difference within the two regions.

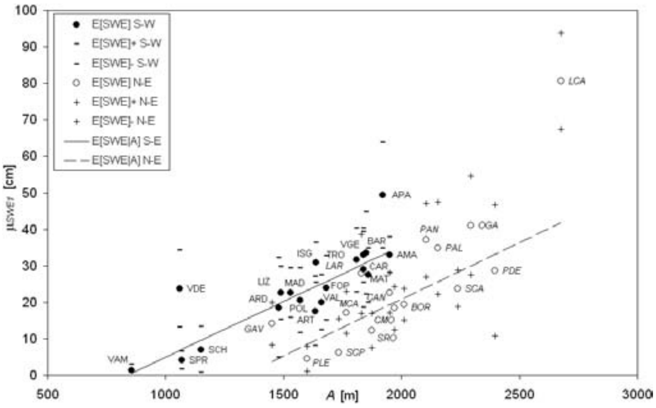

FIG. 2 - Estimated mean value of SWE, at April 1st μ_{SWE1} (E[SWE] in figure legend) against altitude *A* (E[SWE|A] in figure legend) for the two regions S-W and N-E.

Again, these findings are confirmed by the personnel of the AINEVA agency in Bormio based on personal experience, because snowfall upon the northern side of the Orobie alps and Valtellina is known to be lower than its counterpart in the south. This is linked to the main precipitation mechanism given here by the intrusion of moisture laden air masses entering the area moving from S-W to N-E, undergoing uplifting by the relief. By slow degree, water is released through precipitation and less and less moisture is available. This result into a smaller amount of snowfall in the N-E eastern part of the area, *i.e.* «Alta Valtellina». Similar findings were also found by Bocchiola & Rosso (2007b), who showed that H_{72} is considerably higher in region S-W than in region N-E, but with a similar rate of increase against altitude (compare fig. 6 and tab. 2 therein). FRS was also carried out for σ_{SWE1}, giving

$$\sigma_{\mathrm{SWE1}}^{\mathrm{S\text{-}W}} = -8.15 + 0.014 \cdot A, \tag{9}$$

for S-W, while for N-E one has

$$\sigma_{\mathrm{SWE1}}^{\mathrm{N\text{-}E}} = -17.95 + 0.016 \cdot A, \tag{10}$$

with $R^2 = 0.78$ and $R^2 = 0.70$, respectively (shown in fig. 3). It is clear from Eq. (7) to Eq. (10) that the variable actually influencing SWE_1 is altitude. Full summary of the regressions is reported in tab. 2.

DISTRIBUTION OF SWE_1^*

I here investigate distribution of SWE_1^* for both regions S-W and N-E. Their sample frequency, obtained using

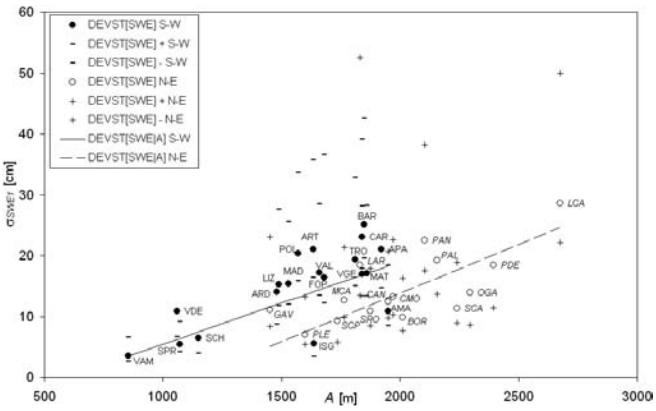

FIG. 3 - Estimated standard deviation of SWE at April 1st μ_{SWE1} (DEVST [SWE] in figure legend) against altitude *A* (DEVST[SWE|A] in figure legend) for the two regions S-W and N-E. Missing upper or lower boundary of confidence interval indicate unreasonable (e.g. negative or unreasonably high) confidence values, due to scarce sample size.

APL plotting position, $F = (i\text{-}0.35)/n_t$, with i rank of SWE_1^* as ordered decreasingly and n_t total sample size, is reported in fig. 4, and compared against a standard normal distribution N(0;1). It is clearly shown there that SWE_1^* is reasonably accommodated by a N(0;1) and is similarly distributed in both regions. This implies that while mean and absolute variability (*i.e.* variance) of SWE_1 do display regional differences, SWE_1^* seems instead homogeneously distributed, thus probably mirroring seasonal behavior of snowpack, affecting the whole region.

ANALYSIS OF SPATIAL COVARIANCE OF SWE_1 INCLUDING GEO-MORPHOLOGIC ATTRIBUTES

To carry out *SWE* estimation at unmeasured sites, one needs regularization of between-station correlation against geomorphologic features (*e.g.* Davis, 1973, Cressie, 1993; Carrol & Cressie, 1997). Here, I calculated the pair wise

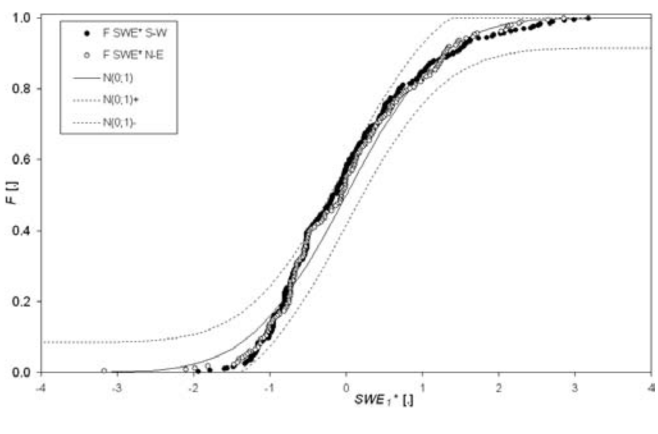

FIG. 4 - Distribution of standardized value of SWE_1, SWE_1* ([SWE* in figure legend]) for the two regions S-W and N-E, and accommodation with a N(0;1) distribution (?=5%).

covariances of the observed SWE_1 values at the different snow gauges $\rho_{i,j} = COV_{i,j}/\sigma_i\sigma_j$. Then, I modeled $\rho_{i,j}$ at gauged sites. A preliminary analysis indicated that sample values of $\rho_{i,j}$ do not display considerable difference between the two regions S-W and N-E. This is shown in fig. 5, where $\rho_{i,j}$ is reported for three cases, namely S-W, N-E and pooled sample, against altitude jump DA. By comparison of fig. 4 with fig. 5 one may infer that behavior of SWE_1^* is likely to be similar within the two regions, in term of both frequency distribution and correlation. Therefore, kriging of SWE_1^* may be carried out for the whole region, while regional issues need to be taken into account in estimation of unstandardized SWE_1, requiring use of its mean and standard deviation.

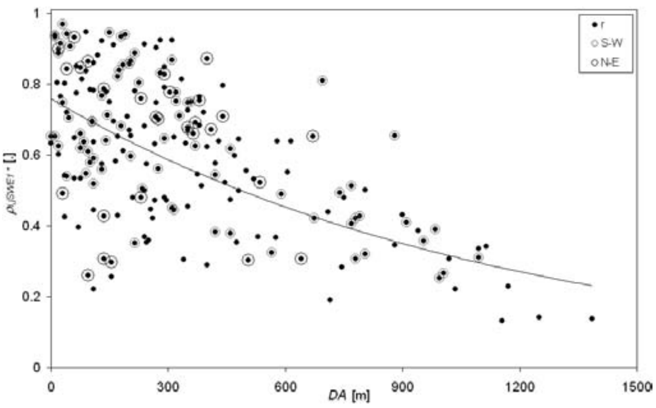

FIG. 5 - Estimated correlation coefficients of SWE_1* between pairs of stations, $\rho_{i,jSWE1*}$ (r in figure legend) against altitude jump DA.

Also, notice that use of the pooled sample, as in fig. 5, provides increased sample size for covariance model estimation. Again here, FSR is carried out to point out the variables most influencing $\rho_{i,j}$. Different models were tested for correlation structure (as in *e.g.* Carrol & Cressie, 1997), namely, linear model, exponential model and quadratic exponential model, against distance, aspect difference, slope difference and altitude jump. However, the best results were provided by an exponential model of $\rho_{i,j}$ depending upon sole altitude jump DA, in meters

$$\rho_{\mathrm{i,j}SWE1*} = 0.759 \cdot \exp\left(-0.0009 \cdot |DA|\right), \tag{11}$$

with determination coefficient $R^2 = 0.29$. This seems quite low. As a comparison, one can consider here the study by Carrol and Cressie (1997). They linked sample $\rho_{i,jSWE1}$ to geomorphic features within North Fork Clearwater river basin (Northern Idaho), using 33-years data from a network of 12 snow gauges covering an area of about 7500 km^2 (150 by 50 km), which is well comparable with this study in the Italian Alps. Using an exponential model for observed covariance field and introducing up to 5 variables (as here, distance, altitude jump, slope difference, aspect difference, plus vegetation cover, not used here after a preliminary analysis), they found a value of residual mean squared error REMSE = 0.11 (Carrol and Cressie, 1997, tab. 2, p. 50, no R^2 reported). For the region here investigated one has REMSE = 0.03, considerably lower, indicating a better fitting of the scatter plot. Therefore, the result here seems to indicate a comparatively good interpretation of the correlation structure of SWE_1 in space. Here however, the only significant variable is altitude, and stations at similar altitudes are considerably correlated, in spite of their distance, which is not significant in this analysis (also, preliminary eyeball assessment showed little, if any, dependence of $\rho_{i,jSWE1*}$ upon distance, not shown for shortness).

CROSS VALIDATION OF SWE$_{10}^*$

To test the accuracy of the kriging approach, cross-validation is carried out here (see *e.g.* Davis, 1973, Carrol & Cressie, 1997). For every station, SWE_1^* for each year in the investigated period is taken as unknown and back estimated by kriging, using the remaining stations where estimated SWE_1 is available, and withholding the unknown station. The covariance matrices in the unknown station *?* and *c* to be used in Eq. (4) and Eq. (5) are estimated depending upon DA, using Eq. (11). Then, SWE_{10}^* at the withheld station is calculated from SWE_1^* values by Eq. (3). Due to the different number of available stations, for each year new matrices Σ and c need to be calculated for each station. Considering the 20 available years (without year 2001) and the 34 stations, a number of 680 simulations were carried out, including estimation of Σ and c, estimation of weights according to Eq. (4), estimation of SWE_{10*} according to Eq. (3) and eventually evaluation of SWE_{10} estimation variance σ_{SWE01}^{*2} using Eq. (5). This required indeed few seconds (the simulations were carried out using an *ad hoc* developed program within Matlab7® environment and a portable PC with Centrino 1.5 Ghz processor and 3 Gb RAM). The standardized estimation error of SWE_1^* for each year y at each station s is calculated

$$\varepsilon^*_{y,s} = \frac{\left(S\hat{W}E^*_{01,y,s} - SWE^*_{01,y,s}\right)}{\sigma^{*2}_{SWE01,y,s}} = N(0;1), \tag{12}$$

which, according to the definition of kriging, should be distributed as a N(0;1). This hypothesis is graphically tested in figs. 6 and 7, where the confidence interval (at a 5% level) for the sample mean and variance of ϵ^* are reported, μ_{e^*} and $\sigma_{e^*}^2$. The confidence intervals are taken

$$\hat{\mu}_{\varepsilon*} - 1.96\frac{\hat{\sigma}_{\varepsilon*}}{\sqrt{n_s}} \le 0 \le \hat{\mu}_{\varepsilon*} + 1.96\frac{\hat{\sigma}_{\varepsilon*}}{\sqrt{n_s}}, \tag{13}$$

with n_s number of data (*i.e.* available years) at the station, for sample mean, and

$$\frac{\hat{\sigma}^2_{\varepsilon*}}{1 + 1.96\sqrt{2/(n_s - 1)}} \le 1 \le \frac{\hat{\sigma}^2_{\varepsilon*}}{1 - 1.96\sqrt{2/(n_s - 1)}}, \tag{14}$$

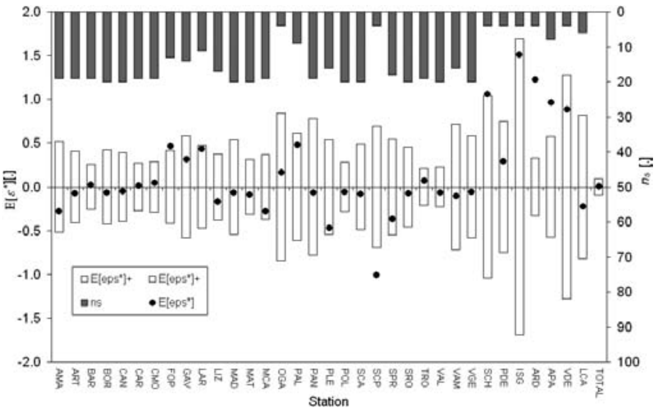

FIG. 6 - Sample mean of standardized estimation error of SWE_1^*, $\mu\Sigma^*$ (E[eps*] in figure legend) and confidence interval (α=5%). Confidence limits given by white boxes. ns is number of years available.

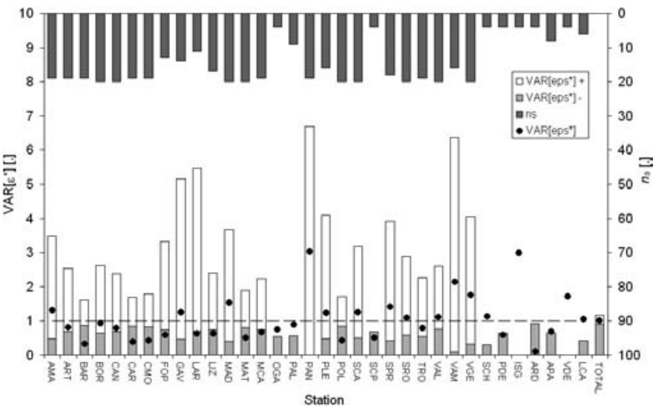

FIG. 7 - Sample variance of σ_Σ^* (VAR[eps*] in figure legend) and confidence interval (α = 5%). Confidence limits given by white boxes. Missing boxes indicate unreasonable (*i.e.* negative or unreasonably high) confidence values, due to scarce sample size.

for sample variance. In few cases the sample values of μ_{ϵ^*} and σ_{ϵ^*} are outside the confidence limits. However, this occurs mainly when a small sample is available (*i.e.* n_s< 10 or so). This is likely due to poor estimation of SWE_1 mean and variance for estimation of SWE_1^*, as well as to poor estimation of μ_{ϵ^*} and $\sigma_{\epsilon^*}^2$. Also, variance $\sigma_{\epsilon^*}^2$ seems somewhat more variable then the mean, as expected for a higher order moment. These findings clearly indicate the need for an increase of sample size n_s for accurate assessment of SWE_{10} using kriging. Notice that the pooled sample (TOTAL in fig. 6 and fig. 7, using 488 values) displays very good matching of the sample statistics of ϵ^* to the unbiasedness (*i.e.* zero mean) and unit variance estimation conditions, supporting the rationale that the greater the available sample the better the results of kriging procedure. Eyeball analysis indicates that the pooled sample of ϵ^* is well accommodated by a N(0;1) distribution, as required (not shown for shortness).

ACCURACY OF SWE_{10} ESTIMATION

In fig. 8, it is reported the average value (within the considered years) of the standard deviation of estimation of SWE_{10}, σ_{SWE10}^2, calculated via Eq. (6) for the 34 snow stations, in absolute value and divided by the local mean value SWE_1 value, μ_{SWE10}, $CV_{SWE10} = \sigma_{SWE10}/\mu_{SWE10}$, against altitude. The latter quantifies the actual amount of error when estimating SWE_{10} using kriging, also with respect to the expected (*i.e.* mean) value in the considered area. While σ_{SWE10} increases with altitude, as expected in view of the increase of SWE_1, the ratio CV_{SWE10} seems pretty constant for A >1000 m asl or so. This seems interesting because most snow cover is at high altitudes, where usually less stations are available. The results herein indicates that the expected error (expressed as a percentage of the mean) is somewhat stationary (and known) for the highest altitudes, so allowing estimation of SWE_1 therein with known accuracy. The accuracy of the kriging technique is also confirmed by the analysis of the determination coefficient for estimation of SWE_{10}, R_{SWE10}^2. This is reported in fig. 9 for each station. It

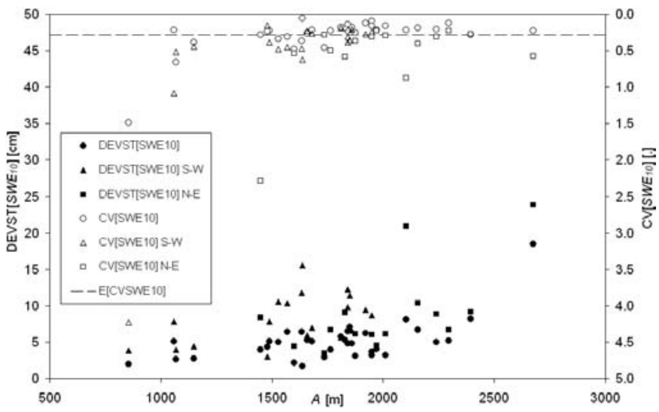

FIG. 8 - Time averaged standard deviation of SWE_{10}, σ_{SWE10}^2 (DEVST [SWE10] in figure legend) absolute and divided by the local mean, $CV_{SWE10} = \sigma_{SWE10}/\mu_{SWE10}$ (CV[SWE10] in figure legend) against altitude.

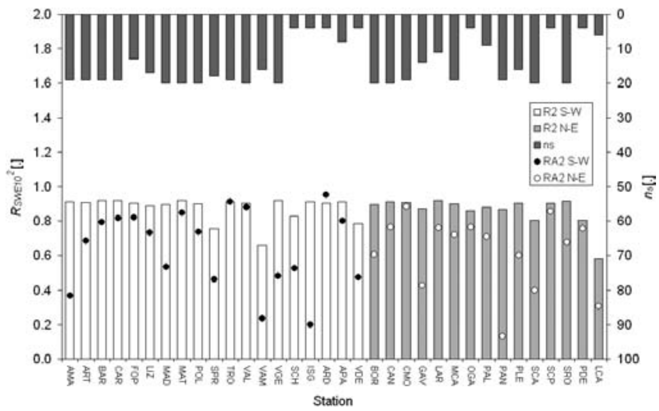

FIG. 9 - Determination coefficient for estimation of SWE_{10}, R_{SWE10}^2 (R2 in figure legend).

is practically constant and its value for the whole sample is $R_{SWE10}^2 = 0.88$, indicating that the kriging procedure allows on the average to capture a considerable part of the variability of SWE_1 distribution in space.

ESTIMATION IN SITES WITH UNKNOWN STATISTICS

When estimating SWE_1 in sites with unknown SWE_1 statistics μ_{SWE1} and σ_{SWE1}, due to scarce or null sample size, one needs to rely upon indirect estimation of the latter. This may result in loss of accuracy of the estimated SWE_1. Here, the effect of indirect estimation of μ_{SWE1} and σ_{SWE1} is tested by back calculating SWE_{10} from SWE_{10}^* using Eq. (2), where μ_{SWE1} and σ_{SWE1} are estimated using Eq. (7) to Eq. (10), yielding SWE_1 statistics depending upon A. Because these equations are different in the two regions, S-W and N-E, these are considered separately. In fig. 8, it is reported the expected standard deviation of estimation of SWE_{10} when μ_{SWE1} and σ_{SWE1} are estimated using A, σ_{SWE10A}, also divided by the local mean value SWE_1 value, μ_{SWE10}, $CV_{SWE10A} = \sigma_{SWE10A}/\mu_{SWE10A}$. Again σ_{SWE10A} increases with altitude and is higher than σ_{SWE10} with known statistics. Also, σ_{SWE10A} is higher for region S-W than for region N-E, as expected in view of the higher variability of SWE_1 therein, as reported in Section 2.5. Instead, CV_{SWE10A} is higher than CV_{SWE10}, and approximately constant for $A > 1000$ m asl or so, and no considerable difference is observed between the two regions. For the whole region one has $CV_{SWE10} = 0.28$, whereas for S-W, $CV_{SWE10A} = 0.59$, and for N-E, $CV_{SWE10A} = 0.54$, indicating in practice a two fold decrease of expected accuracy. In fig. 9 it is reported the determination coefficient of the estimated SWE_1 when its statistics are evaluated based upon altitude, R_{SWE10A}^2, compared against R_{SWE10}^2. The latter is normally higher than the former, and the two regions show in practice similar results (for whole S-W, $R_{SWE10}^2 = 0.88$, and $R_{SWE10A}^2 = 0.65$; for whole N-E, $R_{SWE10}^2 = 0.86$, and $R_{SWE10A}^2 = 0.63$). Eyeball analysis (not reported here) indicates no considerable effect of altitude. These findings clearly indicate that estimation of SWE_1 using kriging in sites with unknown statistics μ_{SWE1} and σ_{SWE1} results into decreased accuracy.

INFLUENCE OF SNOW DENSITY ESTIMATION UPON KRIGING OF SWE

It is possible to evaluate the influence of misestimated snow density ρ_s upon the final SWE estimates. Of the two terms used to estimate SWE, *i.e.* ρ_s and h_s, the former is less variable than the latter. Here, the observed coefficient of variation of snow density (full available data set) is $CV_{\rho_s} = 0.23$, whereas snow depth displays $CV_{hs} = 0.46$, *i.e.* twice as much. The variance of SWE can be calculated starting from mean and variances of ρ_s and h, taken as independent variables, as seen in facts, as

$$\sigma_{SWE}^2 = \sigma_{\rho s}^2\sigma_{hs}^2 + \mu_{hs}^2\sigma_{\rho s}^2 + \mu_{\rho s}^2\sigma_{hs}^2 . \tag{15}$$

Here, ones has an estimated value of $\sigma_{SWE}^2 = 528$ cm^2, against a sample value of $\sigma_{SWE}^2 = 607$ cm^2. Should one use an average constant value of snow density (here, $\mu_{\rho s} = 324$ kgm^{-3}) for SWE estimation (*i.e.* taking null snow density variance, $\sigma_{\rho s}^2 = 0$), from Eq.(15) he would find $\sigma_{SWE}^2(\rho_{const}) = 405$ cm^2. Conversely, taking a constant average depth ($\mu_{hs} = 1.36$ m) would lead to $\sigma_{SWE}^2(h_{const}) = 101$ cm^2. Thus, while use of correct snow depth alone would explain about 77% of the *SWE* variance (*i.e.* 405 cm^2/528 cm^2), use of correct snow density would explain only 19% of this variability (101 cm^2/528 cm^2). Further, because here 51% of snow density is captured, one can modify Eq. (15) to keep this into account and obtain $\sigma_{SWE}^2(\rho_{51\%}) = 467$ cm^2, or 89% of total SWE variability. Therefore, here we can capture almost 90% of the *SWE* variability even if snow density is poorly estimated, in view of its scarce variability, as compared against snow depth. As reported in section 3.2, kriging procedure (based upon the hypothesis of perfectly known SWE_1 values at snow gauges) may show on average an explained variance $R_{SWE10}^2 = 0.88$. One can then guess the expected value of the actual determination coefficient when dealing with uncertain snow density by multiplying R^2 times the percentage of explained variability of SWE_1 using indirectly estimated ρ_s, *i.e.* $R_{SWE10}^2(\rho_{ind}) = R_{SWE10}^2 \cdot 0.9 = 0.78$. Therefore, in spite of uncertain density estimation, kriging procedure still likely provides explanation of about 80% of SWE_1 variability in space. In Section 3.3 it is reported that the effect of unknown *SWE* statistics (*i.e.* of estimation of SWE_1 mean and variance based upon altitude A) results into decreased determination coefficient, *i.e.* $R_{SWE10A}^2 = 0.64$ or so. Therefore, also in case of perfectly known snow density, lack of spatial coverage of the gauging network, which makes local statistics unknown, would have a worse effect that poor knowledge of density.

THE KRIGING METHODOLOGY IN ACTION

To illustrate use of the kriging method, I here show kriged SWE_1 maps for two sample years. These are 1985, with the highest observed values of SWE_1 here (E[SWE_1] = 57.82 cm, evaluated from 25 available measurement stations), and 2003, critical as recalled in the introduction. The average amount of SWE_1 is E[SWE_1] = 10.35 cm, calculated using 23 available measurement stations. To perform kriging for these two years, I used a DTM of the whole region (230-220 m East-North resolution, about 320E^3 cell), and performed calculation of SWE_1^* in each cell (above 500 m asl) using Eq. (3) to (5). For each year, kriging of SWE_1^* required no more than 10 minutes using a portable PC (1.6 Ghz CPU, 3 Gb RAM, Matlab7®). Then, kriged values of SWE_1^* were combined with two maps with mean and standard deviation of SWE_1, μ_{SWE1} and σ_{SWE1}, as calculated from altitude using Eq. (7) to (10). Cells where either μ_{SWE1} or σ_{SWE1} resulted null (or negative) were not included in the calculation. Also, post processing was carried out to exclude negative values of SWE_1 (due to negative values of SWE_1^*, occurring in year 2003, in view of the very low amount of snow). However, these were very few cells, only distributed along the boundary of snowed area. In fig. 10a and 10b, the two resulting maps are shown. Notice the considerable difference in snow amount and snow line with altitude.

The average values of SWE_1 resulting from kriging in 1985 are $E[SWE_{1krige}]$ = 38.99 cm, and $E[SWE_{1krige}]$ = 35.03 cm, for S-W and N-E, respectively. This values are lower than snow gauges sample mean ($E[SWE_1]$ = 57.82 cm), due to presence of snow at low altitudes (above 1000 m asl or so).

In 2003, one has $E[SWE_{1krige}]$ = 14.84 cm and $E[SWE_{1krige}]$ = 16.51 cm, for S-W and N-E respectively, now higher than snow gauges sample mean ($E[SWE_1]$ = 10.35 cm), in view of the highest snow line with respect to year 1985. Also, standard deviation maps of *SWE* could be calculated (not shown for shortness), according to Eq. (6). The average estimated standard deviations in 1985 are DEV.ST $[SWE_{1krige}]$ = 5.90 cm, and 6.37 cm, for S-W and N-E respectively, whereas for 2003 they are DEV.ST$[SWE_{1krige}]$ = 6.13 cm, and 6.49 cm for S-W and N-E respectively.

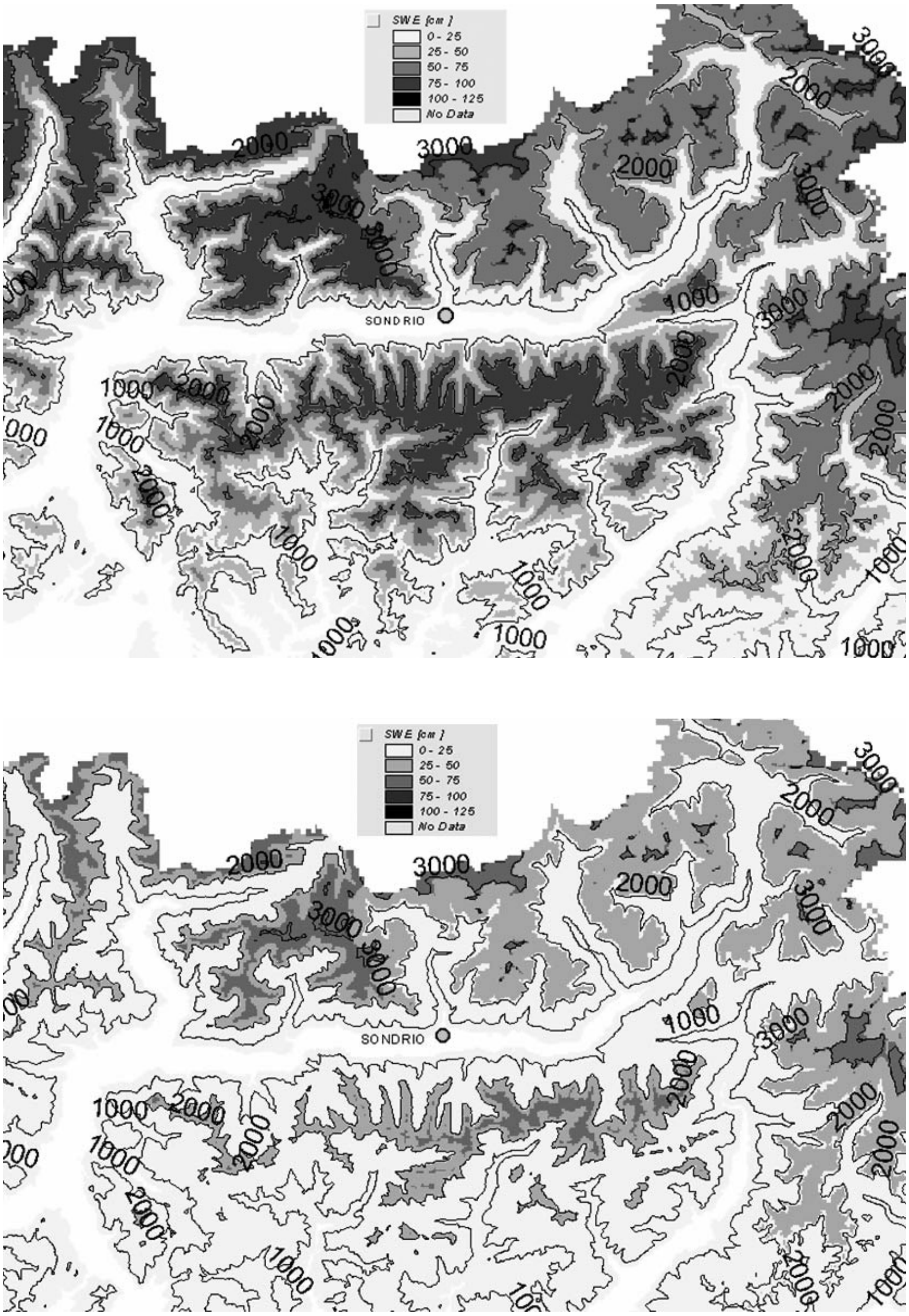

FIG. 10 - Maps of estimated SWE_1 using kriging. a) Year 1985. b) Year 2003.

Slightly increased values for 2003 are likely due to smaller number of available stations (23 instead of 25 in 1985). Comparison with average values of SWE_{1krige} as reported indicates an average uncertainty given by $CV_{SWE1krige}$ = 0.15 and $CV_{SWE1krige}$ = 0.18 for S-W and N-E during 1985, and $CV_{SWE1krige}$ = 0.41 and $CV_{SWE1krige}$ = 0.39 for S-W and N-E during 2003. Clearly, for low amount of SWE_1 kriging error, which is additive in nature, is more relevant.

DISCUSSION AND CONCLUSIONS

From the results here reported, the proposed approach to kriging of SWE_1 in this area provides indeed reliable estimates of known accuracy and can be therefore used accordingly. Whenever SWE_1 statistics are known or can be reasonably well estimated at a given site, the proposed approach allows estimation of the local value at a specific time based upon measurements from the available network, with known accuracy, in the order of 30% of the local average or so. The proposed approach is useful as it provides a relatively simple tool for SWE_1 estimation at ungauged sites, at the cost of knowledge of local altitude and SWE_1 statistics. The kriging procedure is also quick enough, as reported in Section 3.5, that it can be implemented at a watershed scale to evaluate distributed SWE_1, for instance based upon a grid partition. Coupling kriging of available snow measurements with satellite estimation of gridded snow covered area SCA, nowadays widely adopted for water storage assessment in mountain areas (*e.g.* Swamy & Brivio, 1996; Simpson & *alii*, 1998; Cagnati & *alii*, 2004; Hauser & *alii*, 2005; Parajka & Blöschl, 2008) may help in obtaining optimal (in the statistical sense as reported) distributed estimates of SWE_1, say for distributed modeling of snow storage and melting and hydrological and glaciological implications therein. In the future, I will test coupling of kriging with SCA estimation from remote sensing for hydrological purposes in case study watersheds in the investigated region. Some problems arise when accurate *SWE* statistics estimation is unfeasible (compare *e.g.* Carrol & Cressie, 1997). In facts, altitude is the most accurate proxy for *SWE*, and estimation of the latter may be carried by knowledge of the former. This in turn may require dense and evenly spread networks with respect to altitude, and as long as possible data bases, where SWE_1 to altitude relationships may be accurately drawn. Also, such relationships were shown here to be clearly regional in their nature, so that regional investigation is warranted preliminary to setup of kriging procedure. Here snowpack density is rather estimated than measured, so that in the future some adjustment will be necessary to deal with the noise therein. However, the effect of poor density assessment seems smaller than that of lack of measured stations, suggesting that dense spatial coverage is more important than very accurate point snow density evaluation. The present results seem of interest for all those scientist interested in water cycle of mountain areas, dynamics of glaciers, ecology of mountain regions, and prospective impact of climate change thereupon, as well for decision makers willing to pursue knowledge based approaches to water management in the Alps.

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