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Atmospheric precipitable water in Estonia, 1990–2001

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Knowledge on atmospheric precipitable water is necessary as input to hydrological, energetic and radiation models. Short historical review of parameterization of precipitable water is given. For Tallinn (Estonia), simple formulas are proposed to calculate precipitable water from observations of surface water vapor pressure. Seasonal changes of precipitable water in Tallinn are expressed by time series for 1990–2001 as well as by tabulation of monthly averages for this period. Parameterization of precipitable water, decadal time series, and tabulation of monthly averages are also given for three neighboring stations — St. Petersburg (Russia), Jokioinen and Sodankylä (Finland).

Introduction

Precipitable water vapor, or simply, precipitable water, W — the total amount of water vapor in the zenith direction, between the underlying surface (or a surface of certain elevation) and the top of the atmosphere, is an important input to hydrological, energetic and radiation models. Its unit is mass per unit area. In practice this unit is usually considered as the thickness of the layer of liquid water that would be formed if all the vapor in the zenith direction were condensed at the surface of a unit area: 1 mm of the layer corresponds to 1 kg m^{-2}, and 1 cm to 1 g cm^{-1}.

Very little information on the precipitable water regime is available for Estonia. The published information is generalized on the contour maps of monthly average precipitable water above the territory of the former USSR (Institute of Aquatic Problems 1984) and in papers on direct solar beam for neighboring areas (e.g. Abakumova *et al*. 1983, Venäläinen 1994). Therefore, our particular interest was to express precipitable water for Tallinn, Estonia, in terms of surface humidity parameters. The necessity to this kind of parameterization we met calculating attenuation of direct solar beam by different constituents of the atmosphere, including water vapor. Another goal of this work was to study variability of precipitable water, and its mean values, over the region of interest.

As a basic database we used archives of the Tallinn Aerological Station (Estonian Meteorological and Hydrological Institute). In this kind of studies, as usual, low-resolution aerological data were used (Serreze *et al*. 1995). Here, in order to determine whether the use of low-resolution data is justified, we have utilized also high vertical resolution data.

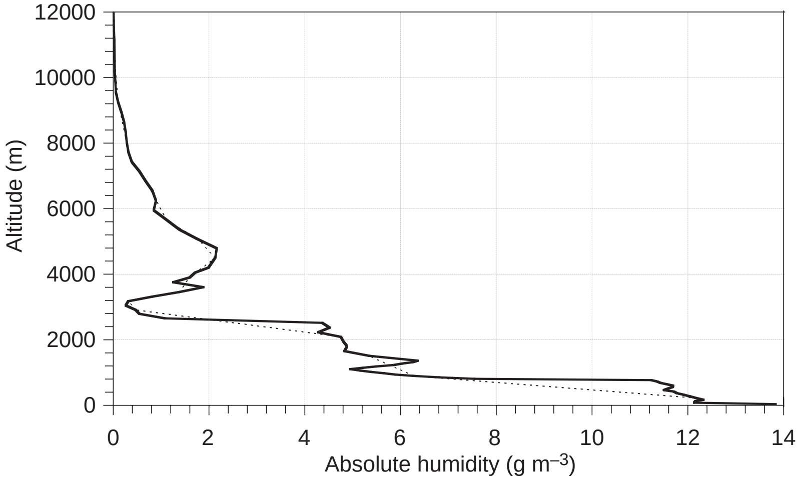

Fig. 1. Vertical humidity profiles in Tallinn on 12 July 1999 at 12 UT. Solid line corresponds to 113 sounding levels, dotted line to 22 sounding levels.

For the sake of comparison and completeness, we derived parameterization of precipitable water, and presented decadal (1990–2001) time series with tabulation of monthly averages also for three neighboring stations in the northern part of the Baltic region — Jokioinen, Sodankylä (Finland) and St. Petersburg (Russia). Three considered stations — Tallinn, Jokioinen, St. Petersburg — have almost the same geographical latitude (between 59 and 61°N), while Sodankylä is located further north (67.36°N), just beyond the polar circle (66.30°N). All these stations belong to the boreal climate region and drainage basin of the Baltic Sea.

Databases

The Tallinn Aerological Station (59.48°N, 24.60°E; 37 m above sea level) is situated at Harku, on the western boundary of Tallinn, Estonia. The station used Vaisala Oy radiosondes of type RS80 since 1993, and sondes RS90 since March 2000. Sondes are launched regularly at 00 and 12 UT.

The whole data set is stored in low vertical resolution according to the WMO TEMP format, consisting of mandatory and significant pressure levels. These low-resolution Tallinn's sounding data for years 1990–2001 are accessible for public use at the Web site of the University of Wyoming (http://www-das.uwyo.edu/upperair/eu.html). Because of the further use of data on precipitable water as input to solar radiation modeling, we focused mainly on daytime sounding profiles starting at 12 UT = 13:38:24 Local Mean Time in Tallinn. It should be mentioned that in the Wyoming's database the year 1996 is only partly covered with daytime profiles and in 1997 observations began in November.

The most recent Tallinn's data (since Dec. 1997) were also used in the high vertical resolution version with up to 400 levels (A. Tõkke, pers. comm.). In order to evaluate the applicability of the limited use of sounding levels for the calculation of W (15–30 levels instead of more than 100), we have compared calculations based on different vertical resolution. It turned out that the average difference was about 2%. This uncertainty is comparable to the error of device (Jauhiainen and Antikainen 1995). An example of this comparison is presented in Fig. 1.

Water vapor density $\rho_v(z)$, at sounding level z, usually expressed in g m^{-3}, also called absolute humidity, is a function of observed absolute temperature T, relative humidity $R_H(z)$ and saturation water vapor pressure $E(z, T)$. The last was calculated using a set of formulas proposed by Hyland and Wexler (1983a, 1983b) and presented on the Honeywell International Inc. Web site (http://content.honeywell.com/). Assimilating water vapor to a perfect gas, its density was obtained by a classic formula, derived from the equation of state of an ideal gas:

$$\rho_v = \frac{217 R_H(z) E(z,T)}{T}, \tag{1}$$

where $E(z, T)$ is in millibars. Integration of Eq. 1 along the vertical profile gives total precipitable water W:

$$W = \int_0^Z \rho_v(z) dz. \tag{2}$$

Concerning Finnish and Russian stations, also only the daytime sounding profiles (12 UT) were adopted from the Web site of the University of Wyoming. Table 1 summarizes the information on locations and used periods. Particular study of daytime clear sky cases was implemented in regard to Tallinn and Jokioinen data. For Tallinn (1990–1996), the NOAA-11 and NOAA-14 satellite images, from the visible (0.58–0.68 μm) and infrared (10.3–11.3 μm) region, were used for visual detection of presence or absence of cloudiness. Named satellite images are available on Web sites of the Dundee Satellite Receiving Station (http://www.sat.dundee.ac.uk/). For Tallinn (1997–2001) and Jokioinen (1996–2000), cloudiness data were adopted from national meteorological archives.

Short historical review

The amount of precipitable water has an evident correlation with several meteorological parameters measured near the underlying surface: partial pressure of water vapor, dew-point temperature, relative humidity, the air temperature, etc. Reviews of parameterizations based on these correlations, are usually presented in connection with studies of attenuation of direct solar beam, e.g. in monographs of Sivkov (1968, 1971) and Iqbal (1983). A short review of parameterization of precipitable water is included also below.

In the first parameterizations (Hann 1906, 1926, Humphreys 1912) precipitable water W (mm) was considered to be proportional to surface water vapor pressure e_0 (mm Hg):

$$W = ae_0, \tag{3}$$

where coefficient a had values 2.0–2.3 for plain locations and lower values for mountain locations (Fowle 1913). Generalizing observations in mountains, Gates proposed an exponential lapse rate of coefficient a (1962, cited by Kambezidis *et al.* 1993):

$$W = 2.3 e_0 10^{-H(\mathrm{station})/22000}, \tag{4}$$

where H(station) represents station's elevations in metres. However, Sivkov substantiated by observations in Karadag (Crimea) during 70 days with low aerosol turbidity, found that linear relation (3) gives too small values in cases of winter weather when temperature inversions often take place. If this is indeed so, the coefficients a, obtained for small e_0 should be higher than for large e_0 , which leads to a nonlinear relationship between W and e_0:

$$W = a(e_0)^n, \tag{5}$$

where $n < 1$. In the Sivkov's study $a = 3.8$ and $n = 2/3$ (Sivkov 1968, 1971).

Method of Russak (1987) represents actually a modification of Eq. 3 considering that coefficients a have different values for each month:

$$a_i = \frac{\overline{W}_i}{\overline{e}_{i0}}, i = 1,...,12, \tag{6}$$

Table 1. Information on locations of sounding stations and used periods.

Station	Latitude	Longitude	Elevation (m)	Period
Tallinn	59.48°N	24.60°E	37	1990–2001
St. Petersburg, Voejkovo	59.98°N	30.60°E	70	1990–1997
St. Petersburg, Voejkovo	59.95°N	30.70°E	78	1998–2001
Jokioinen	60.80°N	23.48°E	103	1990–2001
Sodankylä	67.36°N	26.65°E	179	1990–2001

where $\overline{W}_i$ and $\overline{e}_{i0}$ are monthly means of precipitable water and water vapor pressure for a given location.

Evnevich (1967) also used a linear relationship between W and e_0 but, in order to increase lower values of W, she added a free, constant term b:

$$W = ae_0 + b. \tag{7}$$

Analyzing cities' impact to atmospheric transparency, coefficients a and b were calculated by Abakumova *et al.* (1983) for 25 locations of the former USSR. For example, for Moscow, expressing W in mm, a = 1.6 for all seasons, b = 1.6 for summer, and b = 2.3 for winter.

Reitan (1963) studied monthly means of precipitable water W and station dew-point temperature t_d for 15 stations distributed over the continental United States. He found an excellent linear relationship between ln W (cm) and t_d (°F):

$$\ln W(\text{cm}) = a + bt_d(°\text{F}), \tag{8}$$

where a = –1.466 – (–0.846) and b = 0.0321–0.0476. In the average

$$\ln W\,(\text{cm}) = -\,0.981 + 0.0341t_d\,(°\text{F}), \tag{9}$$

$$\ln W\,(\text{cm}) = 0.1102 + 0.06138t_d\,(°\text{C}). \tag{10}$$

From the fact that precipitable water can be well estimated from surface dew-point temperature, Reitan concluded that moisture in the atmosphere should have a uniform lapse rate, e.g. if the average decrease of absolute humidity is exponential:

$$\rho_v(z) = \rho_v(0)e^{-\beta z}, \tag{11}$$

then, according to Eq. 2, vertical integration from z = 0 until z = Z gives

$$W = \frac{\rho_v(0)}{\beta}\left(1 - e^{-\beta Z}\right). \tag{12}$$

Assuming the moisture to be located only in the layer Z = 9 km, Reitan evaluated that β = 0.35–0.45 km^{-1} and that the most representative value is β = 0.44 km^{-1}. For practical purposes and for further comparisons, it is reasonable to change presentation of the Reitan's result (12). Introducing $1/\beta = H$ = 2272 m and taking Z = 9000 m and $\rho_v(0)$ in g m^{-3}, Eq. 12 rewrites:

$$W\,(\text{mm}) = 0.981H\rho_v(0), \tag{13}$$

here the product "0.981H = 2230 m" may be physically interpreted as depth of the hypothetic vertically homogeneous layer of the atmosphere with the same precipitable water content as the "Reitan's atmosphere". Combination of Eq. 13 with the equation of state (1) gives:

$$W\,(\text{mm}) = 483e_0T^{-1}, \tag{14}$$

where e_0 is in mbars and T in K.

For all Canadian seasons, Won (1977, cited by Iqbal 1983) has developed a simple correlation, similar to the Reitan's one

$$W\,(\text{cm}) = 0.1\exp[(2.2572 + 0.05454t_d\,(°\text{C})], \tag{15}$$

or

$$\ln W\,(\text{cm}) = -1.015 + 0.0303t_d\,(°\text{F}). \tag{16}$$

Parameterizations of Reitan and Won intercept at t_d = –22.8 °C. For warmer temperatures Reitan's model gives higher values of W, e.g., for t_d = 10 °C the difference is 20%.

Leckner (1978) also assumed exponential decay of the water vapor density in the form (11) taking β = 0.439 km^{-1}. Apparently Leckner was not familiar with the Reitan's work. Integration of Leckner's model from z = 0 to z = ∞ gives

$$W\,(\text{mm}) = H\rho_v(0), \tag{17}$$

where coefficient $H = 1/\beta$ = 2278 m represents depth of an equivalent homogeneous layer of humidity in the "Leckner's atmosphere". Expressing e_0 in mbars and applying the equation of state (1), we obtain:

$$W\,(\text{mm}) = 493e_0T^{-1}, \tag{18}$$

this is 2.07% more in comparison with the Reitan's model (14).

Gueymard (1994) essentially perfected approach of Reitan and Leckner. Plotting humidity profiles for different seasons and climatic regions, he demonstrated that exponential humid-

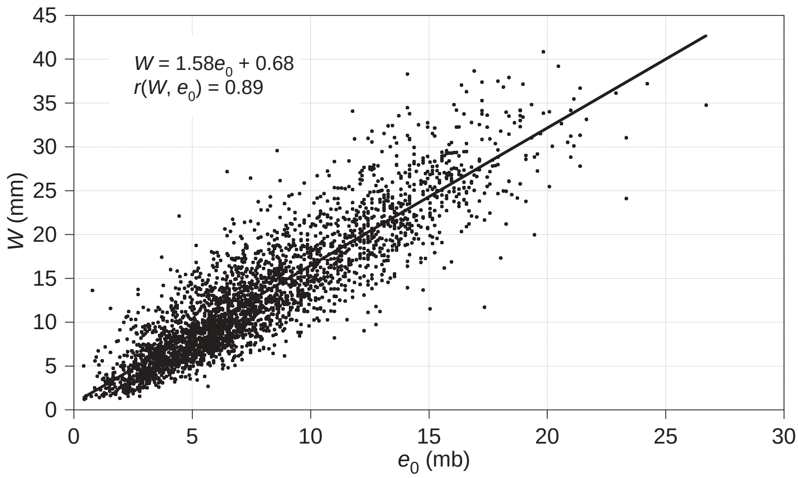

Fig. 2. Precipitable water W (mm) versus surface partial pressure of water vapor e_0 (mb), for all available 12 UT profiles in Tallinn in 1990–2001, 2989 observations.

ity profile (11) agrees rather well to midlatitude summer and subarctic summer. Departure from the exponential profile is obvious for colder atmospheres. The lower the surface temperature, the more frequent become the inversion layers aloft, thus preventing the proper vertical mixing of humidity. The presence of an inversion layer evidently corresponds to an increase of the depth H of an equivalent homogeneous layer of humidity, called by Gueymard as "the apparent water vapor scale height". He showed that, for 21 Canadian and U.S. stations, the average H was about 2.2 km, provided that the surface-level temperature $T > 270$ K. The height H increased sharply at lower temperatures. Gueymard succeeded in expressing H solely as a function of T:

$$H = 0.4976 + 1.52650\theta + \exp(13.68970\theta - 14.9188\theta^3), \quad (19)$$

where $\theta = T/T_0$ and $T_0 = 273.15$ K.

Kondratyev and Moskalenko (Houghton 1984) have developed a simple correlation

$$W\,(\text{mm}) = 14\exp[0.07(T - 288)], \quad (20)$$

where T is the surface-level temperature. It seems that this approximation considerably underestimates content of precipitable water. If the station level relative humidity R_H is 100%, and $T_d = 283.15$ K, this correlation gives only $W = 10$ mm, which is 65% lower of the Won's result.

Parameterization of precipitable water for the region of interest

Figure 2 presents a scatter plot of single values of atmospheric precipitable water W (cm) versus surface partial pressure of water vapor e_0 (millibars) for all available days (clear and cloudy) in Tallinn during 1990–2001 at 12 UT (in total 2989 observations). Minimal values of precipitable water, $W \approx 0.2$ cm, correspond to extremely dry winter atmosphere, maximal values, almost 4 cm, to humid summer air.

Coefficient of cross-correlation $r(W, e_0) = 0.89$, between values of W and partial pressure of surface water vapor e_0, indicates a tense connection between these terms. Can we expect better correlation between W and any other surface-level meteorological parameter as a predictor variable? Or involving several surface-level parameters simultaneously? Using multiple regression analysis, we immersed e_0, T_0, t_d, E, as well logarithm, exponential, etc functions of these parameters (e.g. log e_0, $\exp(e_0)$). Multiple regression, in comparison with the linear one, improved prediction of W only within tenths per cent.

Variation of calculated by selected models values of precipitable water W_{model} around the measured by radiosonde values W_{sonde} we evaluated using the Root Mean Square Error (RMSE; Iqbal, 1983):

$$\text{RMSE} = \left[\frac{\sum_{1}^{N}\left(W_{\text{model}} - W_{\text{sonde}}\right)^2}{N}\right]^{1/2}, \quad (21)$$

where, for Tallinn, $N = 2989$. Beside of described linear model we calculated the RMSE for models of Reitan, Won, Leckner and Gueymard (Table 2). For Tallinn data, nevertheless, simple linear approximation between W and e_0, expressed above by Eq. 7, appeared to be the best fit between precipitable water and surface humidity parameters. Suitability of Eq. 7 in the form of linear regression (Fig. 2):

$$W\,(\mathrm{mm}) = 1.58e_0 + 0.68 \tag{22}$$

is characterized by RMSE = 3.34 mm. According to Table 2, models of Reitan, Gueymard, Leckner and Won secured almost the same accuracy (the RMSE difference is less than 1 mm) and therefore they may be also recommended for approximate calculation on precipitable water in Estonia, particularly in Tallinn. Considering only clear sky cases, we obtained a linear regression:

$$W\,(\mathrm{mm}) = 1.48e_0 + 0.40. \tag{23}$$

The values of precipitable water at clear sky conditions are evidently smaller, compared with cloudy sky conditions. In Tallinn, for 1990–2001, the average values of W at clear sky, calculated by Eq. 23, form 91%–93% of the total means (clear + cloudy) expressed by Eq. 22.

This ratio apparently depends on the dominant type of cloud cover, the profile of temperature, synoptic activity, geographical latitude, etc. According to Hoyt (1978) and Iqbal (1983), precipitable water on a cloudless day is 81% of the mean USA conditions (clear + cloudy sky cases).

Parameterization formula for midnight (00 UT) precipitable water in Tallinn

$$W\,(\mathrm{mm}) = 1.67e_0 + 0.49 \tag{24}$$

gives 3–5% higher values in comparison with the midday (12 UT) values. Considering both, 00 UT and 12 UT soundings in Tallinn during 1990–2001, the parameterization is

$$W\,(\mathrm{mm}) = 1.62e_0 + 0.70, \tag{25}$$

which means 2.5% higher values in comparison with midday. Summary of parameterization formulas for all four Baltic sites is given in Table 3.

Figure 3 presents an example of use of parameterization for calculation of monthly averages of precipitable water for years 1999 and 2000. Solid line is calculated directly from radiosondes' data, and dotted line indirectly, using surface water vapor pressure e_0 and parameterization (Eq. 22) for each sounding. For these two particular years, maximal differences between the monthly averages obtained

Table 2. Variation of modeled values of precipitable water around the observed values in Tallinn, 1990–2001, RMSE — the root mean square error.

Method	RMSE (mm)
This work, Eq. 22	3.34
Reitan, Eqs. 13 and 14	3.36
Gueymard, Eq. 19	3.37
Leckner, Eqs. 17 and 18	3.41
Won, Eq. 15	4.16

Table 3. Parameterization of precipitable water — W, cross-correlation with surface water vapor pressure — $r(W, e_0)$, the root mean square error — RMSE.

Station	W (mm)	$r(W, e_0)$	RMSE (mm)
Tallinn (12 UT, 1990–2001)	$W = 1.58e_0 + 0.68$	0.89	3.34
St. Petersburg (12 UT, 1990–2001)	$W = 1.55e_0 + 1.46$	0.93	2.86
Jokioinen (12 UT, 1990–2001)	$W = 1.58e_0 - 0.12$	0.92	2.76
Sodankylä (12 UT, 1990–2001)	$W = 1.64e_0 + 1.13$	0.93	2.37
Tallinn (clear sky, 1990–2001)	$W = 1.48e_0 + 0.40$	0.91	2.97
Tallinn (00 UT, 1990–2001)	$W = 1.67e_0 + 0.49$	0.89	3.51
Tallinn (00 and 12 UT, 1990–2001)	$W = 1.62e_0 + 0.70$	0.88	3.67
Jokioinen (clear sky, 1996–2000)	$W = 1.54e_0 - 0.44$	0.93	2.60

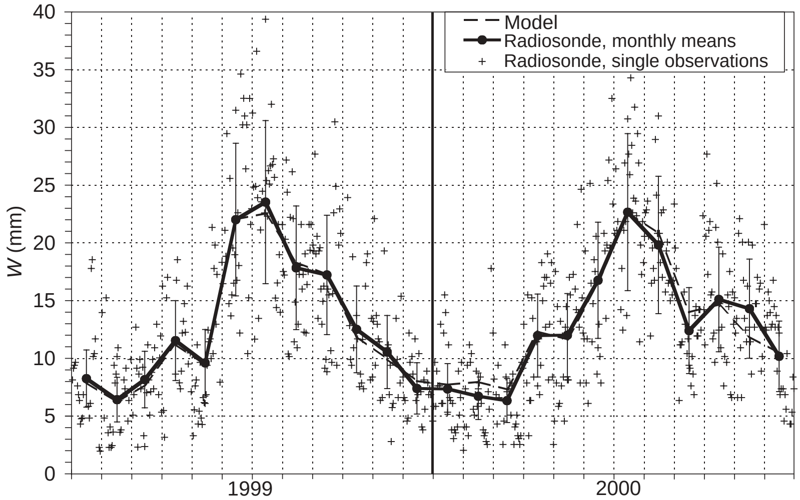

Fig. 3. Time series of monthly average and single values of precipitable water *W* in Tallinn according to all 12 UT soundings in 1999–2000. Solid line — monthly average of radiosonde data, dotted line — monthly average of indirect calculations using parameterization (Eq. 22). Crosses (625 points) represent single sonde observations, vertical bars — their standard deviations calculated for each month.

respectively by direct and indirect method, did not exceed 2.3 mm. The average difference was only 0.6 mm of liquid water. Consequently, the linear parameterization (Eq. 22) is adequate to represent monthly average variability of precipitable water in Tallinn. Figure 3 also contains, in order to visualize scatter of single values of precipitable water, all single daytime sonde observations in Tallinn during 1999–2000 (625 points). The vertical bars show the range of their standard deviations for each considered month.

The cross-correlation coefficient $r(W, e_0)$ for St. Petersburg, Jokioinen and Sodankylä, having values 0.92–0.93, is even higher than for Tallinn where $r(W, e_0) = 0.89$ (Table 3). Values of the regression coefficient, $a = 1.55$–1.64, are in good agreement with results of Abakumova *et al*. (1983). She obtained, according to soundings during 1968–1970 in Moscow, Leningrad, Kiev and Riga, that $a = 1.5$–1.7. Apparently, in order to achieve better correlations (second approximation), than provided by Table 3, seasonal variation of coefficients a and b (Eq. 7) is necessary.

Relationship between single simultaneous observations of precipitable water W in Tallinn and in Jokioinen (160 km from Tallinn), and between Tallinn and St. Petersburg (346 km) was studied. Coefficients of cross-correlation and the Root Mean Square Differences (RMSD, calculated similarly to Eq. 21 for 12 UT observations during 1990–2001 were as follows:

$$r(W_{\text{Tallinn}}, W_{\text{Jokioinen}}) = 0.90, \quad \text{RMSD} = 3.35\ \text{mm}, \tag{26}$$

$$r(W_{\text{Tallinn}}, W_{\text{St Ptrsb}}) = 0.81, \quad \text{RMSD} = 4.57\ \text{mm}. \tag{27}$$

Combination of Eq. 26 with Table 3 gives

$$r(W_{\text{Tallinn}}, W_{\text{Jokioinen}}) \approx r(W_{\text{Tallinn}}, e_0). \tag{28}$$

Let us consider now random fields of atmospheric moisture to be isotropic, with a monotonous correlation function. It follows from Eq. 28 that for sites, located ≈ 160 km from Tallinn, the amount of precipitable water W may be evaluated either by adopting observations in Tallinn or using parameterization (Eq. 22). For longer distances than 160 km, parameterization provides better correlation and lower RMSD than adoption.

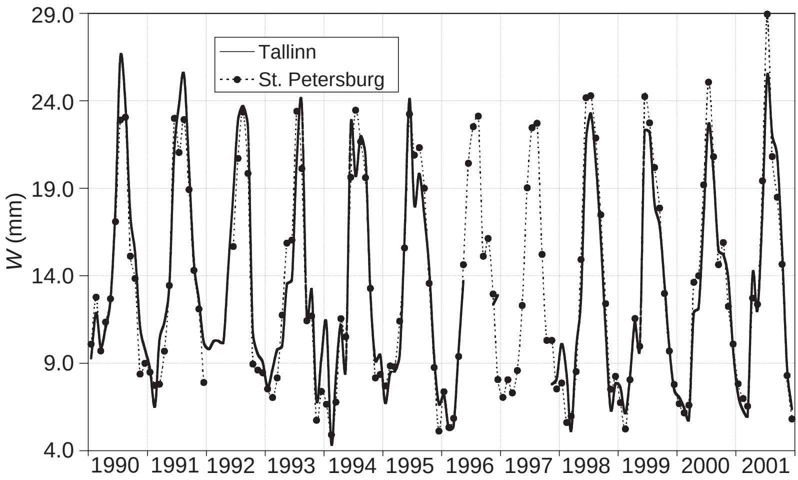

Fig. 4. Time series of monthly average daytime (12 UT) precipitable water for Tallinn and St. Petersburg, calculated from sonde observations, 1990–2001.

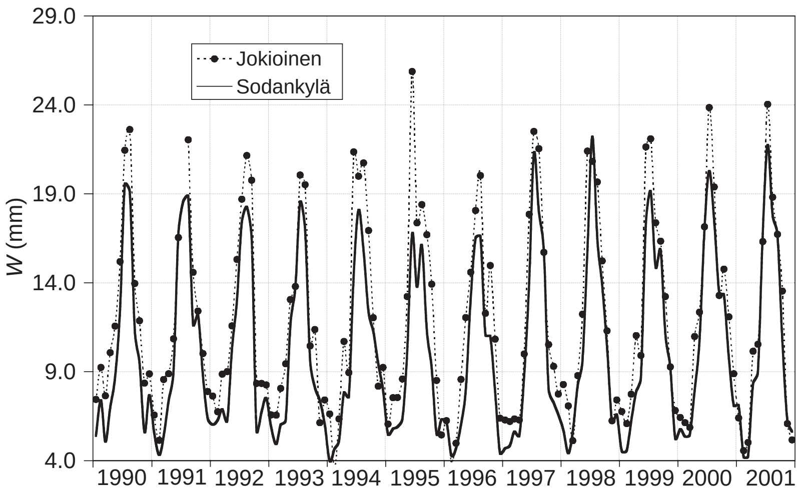

Fig. 5. Time series of monthly average daytime (12 UT) precipitable water for Jokioinen and Sodankylä, calculated from sonde observations, 1990–2001.

Variability of monthly average precipitable water

More completely seasonal changes of monthly average precipitable water W_{month} for all considered sites for period 1990–2001 are presented as time series in Figs. 4 and 5, and as tabulation of monthly means over this period (Table 4). These results are calculated directly from sonde observations, not by means of parameterization. Monthly average precipitable water reaches its minimal values during the cold months (December–February) when usually W_{month} = 5–8 mm. Exceptional was a warm winter 1991/1992 when

Table 4. Monthly average daytime (12 UT) precipitable water *W* (mm), 1990–2001.

Station	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Mean
Tallinn	8.4	7.5	8.0	10.9	12.6	19.7	22.5	21.9	17.9	13.7	9.8	8.9	13.5
St. Ptrsb	7.6	7.2	7.4	10.9	13.3	20.1	23.2	21.8	16.9	13.2	9.4	7.8	13.2
Jokioinen	6.8	6.1	6.8	9.4	11.4	18.1	20.8	20.1	15.2	12.4	8.6	7.5	11.9
Sodankylä	5.5	5.1	5.5	7.1	9.3	15.1	18.9	17.2	13.3	10.1	7.5	6.5	10.1

in Tallinn W_{month} ≈ 10 mm, probably due to high activity of low-pressure systems. It is interesting to note that low values of monthly mean temperature T_{month} are accompanied with low values of precipitable water W_{month} (e.g. February 1994, February 1996). However, low values of precipitable water W_{month} are not always linked with low values of the average temperature T_{month} (e.g. February 2000). Conclusion that temperature is not a good predictor for precipitable water is also supported by low cross-correlation between single values of precipitable water and surface temperature: $r(W, T_s) \approx 0.7$ for Tallinn.

Monthly averages of precipitable water for Tallinn are slightly higher than for St. Petersburg (Table 4). It can be explained by influence on Tallinn's climate of the Baltic Proper (the central part of the Baltic Sea), which is usually ice free during winter months, while the Gulf of Finland near St. Petersburg is ice-covered. During the warmest months (June–August) the average column of precipitable water in Tallinn and St. Petersburg is usually higher than 2 cm and reaches maximal values of W_{month} = 23–27 mm.

Monthly means of precipitable water in Jokioinen and Sodankylä, as in more northern stations, are lower. During winter the lowest values are usually: W_{month} = 4–7 mm and during summer the highest ones: W_{month} = 18–24 mm.

It can be concluded that the summer amount of precipitable water in four considered Baltic sites is about 4 times higher than the winter amount.

Conclusions

Analysis of sounding profiles during 1990–2001 for Tallinn, Estonia, shows that a single value of precipitable water W in Tallinn varies from about $W = 0.2$ to $W = 4$ cm, and monthly average from about $W = 0.5$ to $W = 2.7$ cm. Range of variability of single values of W as well as the monthly averages for St. Petersburg are close to those in Tallinn.

Precipitable water on a cloudless day is 91%–93% of the mean W in Tallinn, calculated for clear + cloudy sky cases.

High cross-correlation (0.89–0.93) between the amount of precipitable water and surface water vapor pressure in four Baltic stations (Tallinn, St. Petersburg, Jokioinen, Sodankylä) was revealed. This high correlation allowed to derive linear regressions (Table 3) for parameterization of precipitable water using surface water vapor pressure. For further study of attenuation of direct solar beam, parameterization formulas for clear sky conditions in Tallinn and in Jokioinen have been developed.

For remote Estonian sites (more than 160 km from Tallinn), parameterization formula (Eq. 22) for calculation of precipitable water provides better results than adoption of Tallinn's observations.

Tabulation of monthly average precipitable water for four named Baltic sites is presented (Table 4).

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