Variability of radiosonde-observed precipitable water in the Baltic region*

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Abstract The total mass of columnar water vapour (precipitable water, *W*) is an important parameter of atmospheric thermodynamic and radiative models. In this work more than 60 000 radiosonde observations from 17 aerological stations in the Baltic region over 14 years, 1989–2002, were used to examine the variability of precipitable water. A table of monthly and annual means of *W* for the stations is given. Seasonal means of *W* are expressed as linear functions of the geographical latitude degree. A linear formula is also derived for parametrisation of precipitable water as a function of two parameters – geographical latitude and surface water vapour pressure.

Keywords Precipitable water vapour; radiosounding; water equivalent

Notation

a, b, c, d	coefficients of a linear formula
e_0	surface partial water vapour pressure (hPa)
$e(z)$	partial water vapour pressure at level z (hPa)
$E(t)$	saturation water vapour pressure for temperature t (hPa)
p	atmospheric pressure (hPa)
R^2	coefficient of determination
R_H	relative humidity
$t, t(z)$	temperature at level z (°C)
U	uncertainty of measurements
W	precipitable water (kg m^{-2})
z	elevation above sea level
φ	geographical (North) latitude degree
λ	geographical (East) longitude degree

Acronyms

ASL	Above Sea Level
GPS	Global Positioning System
Lat N	geographical (North) latitude degree
LMT	Local Mean Time
Lon E	geographical (East) longitude degree
MBE	Mean Bias Error

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MST	Missing Station Test
NAO	North Atlantic Oscillation
STDEV	Standard Deviation
UTC	Coordinated Universal Time

Introduction

The total mass of columnar (integrated) water vapour, usually called precipitable water vapour or simply, precipitable water, *W*, is a fundamental quantity for all atmospheric sciences. The global mass of the atmosphere varies mainly from changes in water vapour loading (Trenberth and Smith 2005). As the most important greenhouse gas, water vapour strongly modulates propagation of solar and terrestrial radiation and plays a crucial role in the Earth's radiative budget. Beside this, water vapour also has a significant influence on the accuracy of satellite monitoring information about surface properties (satellite images) and on GPS applications.

Precipitable water is expressed as mass per unit area. This unit, *mass per unit area*, is in practice usually given as the thickness of the layer of liquid water that would be formed if all the vapour 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^{-2}.

In actinometry, to simplify theoretical analysis, as well as for practical calculations of extinction of the direct solar beam, the atmosphere is considered as a succession of individual attenuation layers (Leckner 1978; Gueymard 1998). At least three layers are necessary: (1) clean and dry or *an ideal atmosphere*, (2) water vapour and (3) aerosol particles. These layers can be considered independent of each other so that the total beam radiation which reaches the ground level is obtained as a direct product of individual transmittances. Within narrow spectral regions, the physical order of attenuation layers is irrelevant. When the attenuation is extended to cases of a broadband solar beam, transmittance of a layer depends on its vertical location.

Total broadband transmittance as a whole (i.e. the product of three individual transmittances) represents one output of regular pyrheliometric observations. Transmittance of an ideal atmosphere depends on Rayleigh scattering, ozone absorption and absorption by uniformly mixed gases such as CO_2, NO_2, etc, and can be calculated with high accuracy. Models which allow calculation of transmittance of the "water vapour layer" use a value of precipitable water as an input parameter. Having observed the value of total transmittance and evaluated transmittances of two layers, an ideal atmosphere and a water vapour, aerosol transmittance remains a residual part of the total transmittance. In such a case, observations of the direct solar beam contain information on air quality in the atmospheric column, provided that transmittance of an ideal atmosphere is calculated and the amount of precipitable water, *W*, is known.

The number of methods for observation of *W* has increased considerably during the last few decades and now includes optical ground-based and satellite soundings, microwave and GPS soundings – modern techniques which enable us to follow quick changes in column water vapour content. However, classic balloon aerologic radio-sounding remains a dominant routine meteorological method. This is the only method for retrospective studies of atmospheric humidity content during previous decades. Due to expensive instrumentation, the network of aerologic stations in the Baltic area is sparse, with a distance between neighbouring stations of 170–370 km. From several stations balloons are launched only once daily. Therefore, especially for solar radiation and aerosol studies and also for correction of satellite images, elaboration of approximate methods to estimate precipitable water is of considerable current interest.

In our previous work (Okulov *et al.* 2002), we gave a short historical review for this kind of parametrisation and demonstrated that, for conditions at Tallinn, a simple linear relationship between W and surface water vapour pressure, e_0, appeared to be the best fit between W and surface humidity parameters. We presented linear parametrisation formulae for four Baltic locations (Tallinn, St Petersburg, Jokioinen, Sodankylä).

In this work we increase the number of considered aerological stations to 17 and give a linear two-parameter approximation of W: against e_0 and the geographical latitude degree, φ. We also express seasonal means of precipitable water as functions only of the geographical latitude. The necessity for better latitudinal and seasonal parametrisation of W in climatological studies is highlighted by the disturbing fact that atmospheric models underestimate the absorption of sunlight (Maurellis and Tennyson 2003). As a probable explanation for "the surplus radiation" in the global energy budget, these authors suggested insufficient representation of water vapour absorption bands, especially in the optical and ultraviolet regions. Another reason, however, may be hidden by an insufficient representation of the planetary distribution of precipitable water.

As a general orientation, we first list some extreme and typical values of precipitable water for different climatic conditions. In Tallinn, the minimal values of W are about 1 mm on cold winter days and about 40 mm on warm and humid summer days; the mean annual value in Tallinn during 1989–2002 was 14.2 mm. For comparison, in Calcutta (22.65°N, 88.45°E), according to a database described below, a maximal value of $W = 94.9$ mm was observed during 16 July 2003.

Atmospheric (optical) models often use fixed values of precipitable water for certain climatic conditions. For example, the MODTRAN models use six values of W: 41.2 mm (tropical), 29.3 mm (midlatitude summer), 20.8 mm (subarctic summer), 14.2 mm (the 1976 US Standard Atmosphere), 8.5 mm (midlatitude winter) and 4.1 mm (subarctic winter) (Qiu 2001). The global water vapour content in the atmosphere was estimated by Peixoto and Oort (1992) to be of the order of 13.1×10^{15} kg and, most recently, Trenberth and Smith (2005) published 12.7×10^{15} kg as a new estimate. These estimates are equivalent to a uniform layer of 25.7 mm and 24.9 mm of liquid water covering the globe, respectively.

The atmospheric humidity regime in the Baltic area depends on, beside seasonal, diurnal and latitudinal variations, the origin of air masses which arrive at this area. During the four months December–March (DJFM), a typical synoptic situation is a huge low-pressure area over Iceland and a high-pressure area over the Azores. This situation favours westerlies which extend as far eastward as central Siberia. Meridional (north–south, south–north) shifts of air masses are weak. The strength of this typical situation is characterised by the North Atlantic Oscillation (NAO) index. The relation between the eastward transport of marine air across the 10°W meridian in the latitudinal belt 30–80°N during DJFM and the NAO was studied by Ruprecht *et al.* (2002). This work revealed that, during the high NAO winters, the maximum of the vertically integrated humidity transport, 160 kg/(m s), is observed between 50–55°N. During the low NAO winters, this maximum reduces to 120 kg/(m s) and shifts to 44°N. The synoptic situation is considerably more complicated from April to November. The meridional component in the movement of air masses is more important and air can reach the Baltic area even from the east. An all-seasons generalisation of synoptic situations in the Baltic region apparently has to rely on circulation patterns and weather types (Sepp and Jaagus 2002; Post *et al.* 2002; Mietus 1998).

Data

Night-time radiosonde reports (00 UTC) from 17 aerologic stations for 1989–2002 in the Baltic area (Table 1, Figure 1) were used to obtain values of W. The northernmost site, the Finnish station at Sodankylä (67.36°N), is located just beyond the polar circle (66.55°N),

Table 1 List of aerological stations in the Baltic region, their monthly and annual nighttime (00 UTC) average radiosonde precipitable water *W* (mm), 1989–2002. The last column gives the ratio of summer (JJA) and winter (DJF) seasonal means

Station	Lat N	Lon E	*z* (m)	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Year	Ratio
Sodankylä	67.36	26.65	179	6.4	5.6	6.1	8.5	10.7	15.5	19.3	18.2	13.9	10.4	7.8	6.9	10.8	2.81
Lulea	65.55	22.13	34	6.7	5.6	6.4	7.9	10.5	16.1	20.8	19.4	14.6	10.8	7.7	6.6	11.1	2.97
Sundsvall	62.53	17.45	6	7.4	6.5	7.6	9.4	12.0	17.8	22.6	21.5	16.7	12.6	8.9	7.6	12.5	2.88
Jyväskylä	62.40	25.68	145	6.7	6.1	7.2	8.9	11.1	16.6	19.6	19.7	14.5	11.4	8.5	6.9	11.4	2.84
Jokioinen	60.81	23.50	103	7.5	7.0	7.4	9.9	11.7	17.6	20.9	20.5	15.7	12.6	9.3	7.5	12.3	2.68
Voejkovo	59.95	30.70	78	8.0	7.9	7.9	11.2	13.7	21.2	24.7	22.6	17.4	13.2	9.7	7.7	13.8	2.90
Tallinn	59.38	24.58	34	8.9	9.2	9.4	12.1	13.7	19.8	23.4	22.3	17.3	14.3	11.0	8.7	14.2	2.44
Göteborg	57.66	12.50	164	8.4	8.6	8.7	10.9	13.2	16.7	20.3	20.8	17.0	13.9	10.3	8.8	13.1	2.25
Visby	57.65	18.35	47	8.7	8.1	8.5	10.8	12.9	17.6	21.5	22.5	17.8	14.1	10.9	8.9	13.5	2.39
Riga	56.96	24.05	26	8.3	8.5	8.6	12.0	14.5	20.9	24.4	24.3	19.2	14.8	10.9	8.6	14.6	2.73
Copenhagen	55.76	12.53	42	10.1	10.0	9.9	11.8	14.4	18.1	22.0	21.9	18.7	15.3	11.5	9.9	14.5	2.07
Leba	54.75	17.53	6	10.1	9.3	9.4	12.2	16.1	20.2	23.7	24.1	20.2	16.2	11.8	9.8	15.2	2.33
Schleswig	54.53	9.55	48	10.4	10.5	10.7	12.3	15.6	19.3	22.6	23.1	20.0	16.1	12.2	10.3	15.3	2.08
Greifswald	54.10	13.40	6	10.5	10.0	10.1	12.9	16.8	19.8	23.4	23.8	20.3	15.8	12.3	10.2	15.5	2.18
Legionowo	52.40	20.96	96	9.5	9.0	9.5	13.6	17.6	22.1	25.9	24.0	19.3	16.2	12.1	9.2	15.7	2.60
Lindenberg	52.21	14.11	115	9.4	9.6	10.2	12.2	17.3	20.9	24.2	24.2	20.3	15.9	12.2	10.4	15.6	2.36
Wroclaw	51.78	16.88	122	10.2	10.7	11.2	13.2	17.9	21.9	25.1	24.5	20.7	17.1	12.2	9.9	16.2	2.32

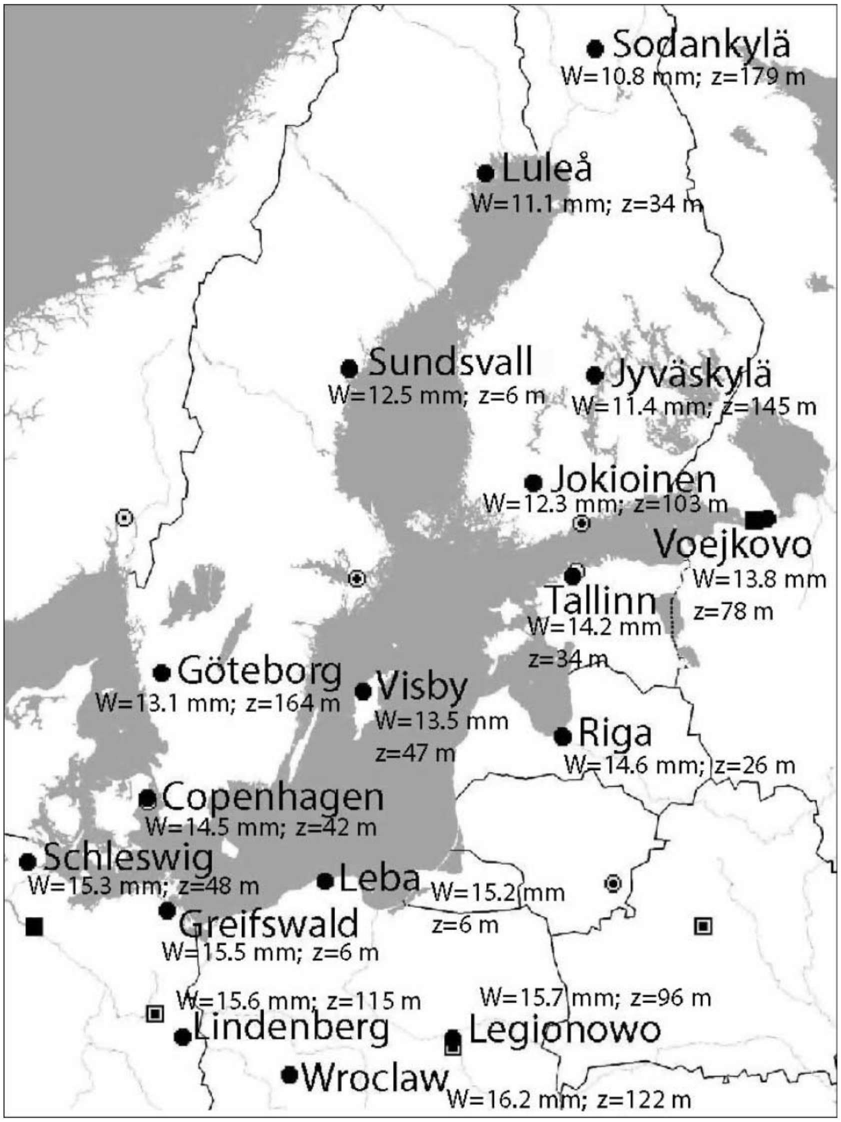

Figure 1 Map of 17 considered aerological stations in the Baltic region; W – the mean annual value of precipitable water (1989–2002), z – the elevation above sea level

while the southernmost location, the Polish station at Wroclaw (51.78°N), is situated 15.58° (1730 km) southward. Actually, there are additional stations in this region (e.g. Pskov), but they were excluded because their time series are too inhomogeneous or short or they do not have observations at 00 UTC.

Usually radiosondes are launched twice daily, at 00 UTC and 12 UTC. We omitted the 12 UTC observations for several reasons. First, and this is the main reason, the set of 12 UTC soundings is less complete. For example, in Tallinn, the 12 UTC observations were stopped after 2001. The practice to confine the analysis to 00 UTC observations has also been applied by Ross *et al.* (2002), who examined radiosonde data for 150 stations in US and Canada.

Second, the diurnal cycle of precipitable water is usually weak compared to the overall value of W. According to dual-channel microwave radiometer observations in Potsdam (52.38°N, 13.07°E, 81 m ASL) during 1–14 August 1997, when W varied within the limits of

25 ± 10 mm, the mean interdiurnal variation (peak-to-peak difference between the extremal values during 24 h) was 2.2 mm, or less than 10% of the daily mean of W (Güldner and Spänkuch 1999).

This particular result was confirmed by a more extensive GPS investigation for a set of 30 sites during 1995–1997 (Bouma and Stoew 2001). Selecting, for our purpose, from this set 26 locations, mainly from Finland and Sweden, which belong to the latitudinal belt 52–68°N, we found the average diurnal peak-to-peak difference to be 0.5–1.5 mm for both the winter (DJF) and the mid-to-late summer (JAS) months. Including again four southern locations (46–49°N), the winter peak-to-peak diurnal difference did not change, but the summer one increased to 2–3.2 mm. During the winter months, the time of the maximum daily value of W for 26 Baltic locations demonstrated a relatively large scatter, from 05 UTC to 20 UTC. But for the summer (JAS) months there was a clear two-hour afternoon period, 15–17 UTC, when the daily maximum of W was reached.

Naturally, choosing a single month at a particular site, the average diurnal scatter of W is smaller than for the 3-month (e.g. JAS) period. For example, in September (1995–1997) in Onsala (57.40°N, 11.92°E, 10 m) the average diurnal peak-to-peak difference was less than 1 mm. The time of the highest daily value in September, $W = 16.8$ mm, was shifted to midday (12 UTC), leaving for 15–17 UTC a 2-h minimum, $W = 16.0$ mm. At midnight (00 UTC) there was a secondary maximum, slightly lower than the principal maximum at 12 UTC.

Radiosondes in Onsala were launched 4 times daily: 00, 06, 12 and 18 UTC. This time resolution does not allow us to follow details of the diurnal cycle of W but it appeared that in September (1995–1997) the midnight value exceeded the midday one by 2 mm: W(00 UTC) = 17 mm, W(12 UTC) = 15 mm. In July the difference was less than 1 mm, but the same tendency was observed.

Discrepancies between diurnal cycles of GPS and radiosonde observations should be solved by further investigations. But the work of Bouma and Stoew proved that in the Baltic region, W(00 UTC) and W(12 UTC), at least for the winter months (DJF) and for mid-to-late summer (JAS), on the average differ by less than 2 mm. For several applications (e.g. the estimation of aerosol optical depth), this knowledge allows us to use nighttime values of W for daytime conditions.

In our work we have used sounding profiles in the low vertical resolution WMO TEMP format, accessible for public use at the website of the University of Wyoming (http://www.uwyo.edu). We considered only soundings with at least 14 levels; usually 15–30 levels were represented. For each sounding, precipitable water is already calculated and presented on the website. Although the calculation scheme is not provided, it appeared that for each level z, using observed relative humidity R_H and temperature $t(z)$, water vapour pressure, $e(z) = R_H E(t)$, was calculated. Then, assuming water vapour to be a perfect gas, its density was found and integrated along the vertical profile. We verified that saturation water vapour pressures $E(t)$, for temperatures less than 0°C, were also calculated with respect to water, in correspondence with WMO recommendations (WMO 1988).

We estimated that the relative standard deviation for precipitable water W(high), calculated from high-resolution radiosonde data (100–500 observational levels), is 4% and for W(low), calculated using low vertical resolution (15–30 levels), is 5%. The relative difference

$$\text{relative difference} = \frac{W(\text{low}) - W(\text{high})}{W(\text{high})} \times 100\% \qquad (1)$$

due to the use of low vertical resolution data in numerical vertical integration is usually less than 2%, thus less than the error of the device. Larger relative differences (10%) appear in cold months, when the values of W are small. The average annual difference between

W(high) and W(low), or the mean bias error, is only 0.06 mm. This justifies the use of low-resolution sounding profiles for calculation of precipitable water.

At a confidence level of 95%, uncertainties for the observed meteorological parameters as measured by the Vaisala RS90 radiosondes are as follows: pressure, $U(p, 95\%) = 1.5$ hPa; temperature, $U(t, 95\%) = 0.5$°C; relative humidity, $U(R_H, 95\%) = 5\%$. Considering these uncertainty values, a relative uncertainty of a radiosonde-observed precipitable water (uncertainty of device) is 7.7%. Combined with the uncertainty due to the use of low-resolution data, the total relative uncertainty for W(low) is 9.3%. For a typical summer single observation of $W(\text{low}) = 20$ mm, this leads to an uncertainty of observation, $U(W, 95\%) = 1.9$ mm (Jakobson 2004).

Seasonal precipitable water *versus* latitude

Long-term (1989–2002) monthly and annual averages of nighttime (00 UTC) precipitable water for each station are presented in Table 1. Within the Baltic region, the long-term monthly mean precipitable water varies from $W = 5.6$ mm (Sodankylä and Luleå, February) to 25.9 mm (Legionowo, July). The ratio between the average summer (JJA) and the average winter (DJF) value, representing the seasonal contrast, tends to decrease in the southwest direction. This can be explained by cyclonic winds from the southwest direction that prevail almost during the year (except April and May) in the Baltic Sea area and carry humid air from the Atlantic Ocean (Mietus 1998).

In order to study the southward increase of precipitable water, we averaged its 00 UTC values over the four seasons: spring (MAM), summer (JJA), autumn (SON) and winter (DJF). Apparently due to the relatively small north–south extent of the Baltic area, a simple linear fit:

$$W(\text{season}) = a\varphi + b \tag{2}$$

was found to express the dependence of seasonal means of W on the latitude degree φ (Figure 2). A linear expression, Equation (2), for each season is given in Figure 2.

Scatter of seasonal means from the linear fit is greatest during summer and winter (coefficient of determination $R^2 = 0.67$ and 0.86, respectively). During these seasons, the proximity of two different environments – oceanic and continental – especially affects the properties of air, resulting in deviation from the mean latitudinal trend. Another factor affecting humidity is the extent of the Baltic Sea in the meridional and zonal directions, which precludes uniform spatial variations of air temperature and humidity in the region (Mietus 1998).

In winter Atlantic airmasses which move over Denmark (the Jutland cyclones) keep water in the southern and middle part of the Baltic Sea unfrozen (Ice Atlas 1982). This explains the slightly (less than 0.8 mm or 8%) higher winter values of W than predicted by the linear latitudinal trend, Equation (2), in Schleswig, Copenhagen, Greifswald, Leba and Tallinn. The influence of oceanic and marine environments is smallest in Legionovo where the predicted latitudinal value overestimates the observed multiannual winter mean by 0.9 mm (10%).

In summer the latitudinal mean underestimates the observed seasonal mean in Voejkovo (by 2.3 mm or 10%) which is the easternmost station. In Göteborg, in contrast, air is cooled by Atlantic winds and the linear trend, Equation (2), overestimates precipitable water by 2.1 mm (11%).

In February the intensity of mean air flow over the Baltic region starts to decrease. In April and May, due to the weakened Icelandic Low, the Atlantic air flow is very weak, the pressure pattern above the area becomes irregular and air moves even from east to west (Mietus 1998). The atmosphere over the region is well mixed and the latitudinal trend describes meridional changes of W(season) with high rate of determination ($R^2 = 0.93$).

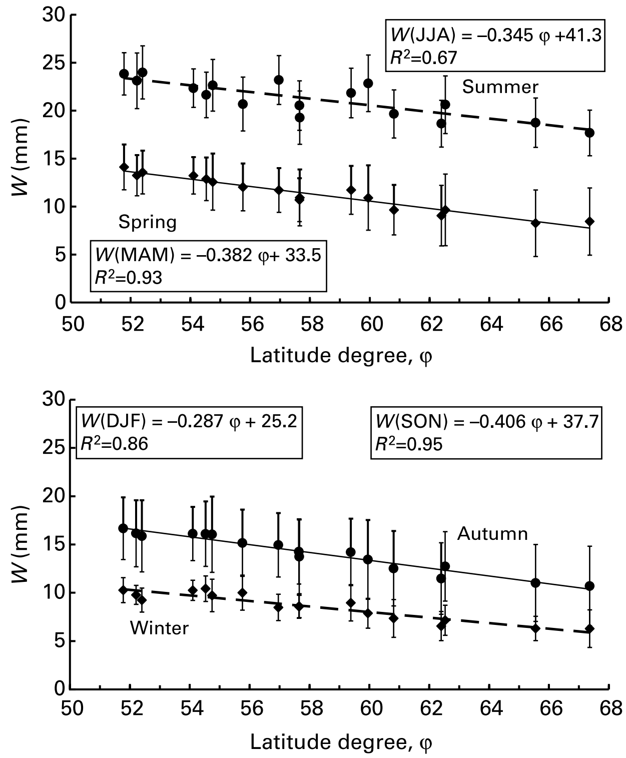

Figure 2 Mean seasonal precipitable water as a function of geographical latitude, φ, 1989–2002. The error bar is the standard deviation between single monthly means and the mean seasonal value. Four linear equations between latitude and seasonal average precipitable water are given. Coefficient of determination, R^2, evaluates applicability of the linear formula

Autumn (SON) in the Baltic area is characterised with strong winds (Mietus 1998). The atmosphere over the region is again well mixed and a linear trend describes meridional changes of *W*(season) with high rate of determination ($R^2 = 0.94$).

We also tried to improve the parametrisation, Equation (2), by including for a given location its elevation above the mean sea level, *z*, and the geographical longitude degree, λ:

$$\mathrm{W(season)} = a\varphi + b + cz \tag{3}$$

$$W(\mathrm{season}) = a\varphi + b + d\lambda \tag{4}$$

but the use of altitude *z* was statistically justified only for autumn, then the coefficient $c = -0.0054$ mm/m. The use of the longitude parameter λ was not statistically justified for our database.

Precipitable water versus surface water vapour pressure and the latitude degree

According to an analysis performed by Okulov *et al.* (2002), a linear expression between *W* and midday (12 UTC) surface water vapour pressure e_0,

$$W = ae_0 + b \tag{5}$$

appeared to be the best fit between precipitable water and surface humidity parameters in Tallinn. We generalised this result for the Baltic Sea region considering coefficients *a* and *b* as linear functions of latitude φ. The coefficients were found by applying the least square method to more than 60 000 soundings from 17 stations during 1989–2002:

$$a = -0.0089\varphi + 2.16 \tag{6}$$

$$b = 0.132\varphi - 7.4 \tag{7}$$

which leads to a two-parameter all-seasons formula for the approximate calculation of single values of W:

$$W = 2.16e_0 - 0.0089e_0\varphi + 0.132\varphi - 7.4 \tag{8}$$

where W is in mm, e_0 is the midnight (00 UTC) surface water vapour pressure in hPa and φ is in degrees.

Uncertainty analysis

In the previous sections we presented two methods for the estimation of precipitable water, W, in the Baltic region: (1) a rather simple one-parameter linear method which, for a given latitude degree, φ, allows us to obtain only four different values of W as its seasonal mean (Equation (2): we shall name this Model A) and (2) a two-parameter linear method for calculation of W according to given surface midnight (00 UTC) water vapour pressure, e_0, and geographical latitude degree, φ (Equation (8): Model B).

Now we analyse the accuracy of these models. We use three different evaluation methods: (1) missing station test (MST), (2) mean bias error (MBE) and (3) standard deviation (STDEV).

Using the MST, we first calculated new parameters (constants) of models using data from 16 stations and then repeated calculations using all 17 stations. It appeared that the parameters of models calculated, either on the bases of 16 stations or 17 stations, were close. Testing Model A, the maximal difference of W (0.2 mm) occurred in the case when the missing station was Legionowo in summer. Testing Model B, we inserted the latitude degree and mean seasonal values of water vapour pressure for the location of the omitted station. The use of two sets of stations (16 or 17) caused the maximal difference of 0.5 mm for the southernmost station, Wroclaw ($e_0 = 13$ hPa, $\varphi = 51.78°$). For both models the missing station test shows that the maximal differences are smaller than the uncertainty of a single observation of W. Thus, applying the MBE and the STDEV tests below, we use both models, A and B, as derived for the initial set of 17 stations.

For both models the MBE and the STDEV were calculated according to the next general formulae:

$$\mathrm{MBE}(W) = \frac{1}{N}\sum_{1}^{N}(W_{\mathrm{model}} - W_{\mathrm{radiosonde}}) \tag{9}$$

$$\mathrm{STDEV}(W) = \sqrt{\frac{1}{N-2}\sum_{1}^{N}(W_{\mathrm{model}} - W_{\mathrm{radiosonde}})^2} \tag{10}$$

where N is the number of observations at a given station during all given seasons in 1989–2002 (e.g. during 14 summers in Tallinn) and $W_{\mathrm{radiosonde}}$ is an observed individual value of precipitable water. The results of both tests are presented in Fig. 3.

The MBE is an indication of the average deviation of the predicted values from the measured ones. Ideally a zero value of MBE should be obtained. Concerning Model A, the MBE exceeds 2 mm only in Voejkovo under summer conditions when the model underestimates the actual precipitable water by 2.1 mm. Model B overestimates W in summer in Göteborg by 2.0 mm and in Copenhagen by 2.3 mm, while in other locations and seasons the MBE is less than 2 mm.

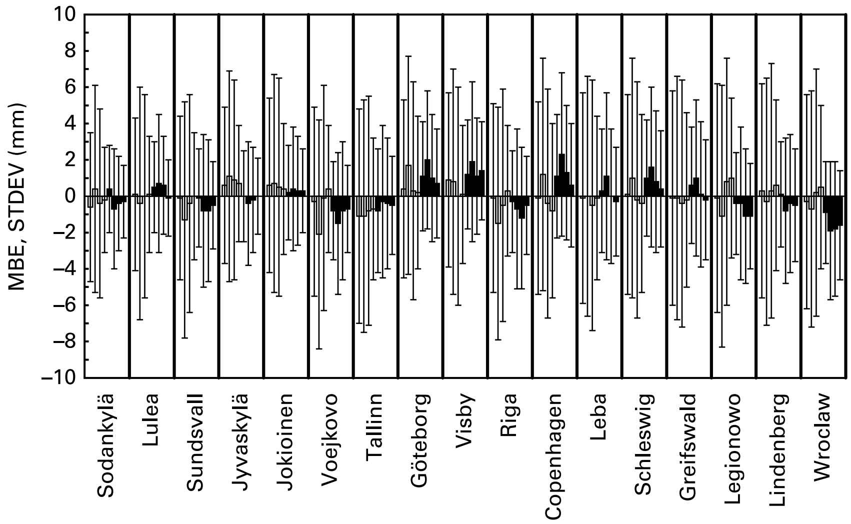

Figure 3 Seasonal mean bias errors (MBE) as differences between the modelled values W_{model} and observed single values of radiosonde precipitable water $W_{\text{radiosonde}}$: Model A, empty boxes, $W_{\text{model}} = W(\varphi, \text{season})$, Model B, filled boxes, $W_{\text{model}} = W(\varphi, e_0)$. A succession of the boxes corresponds to spring (MAM), summer (JJA), autumn (SON) and winter (DJF), respectively. Error bars represent the standard deviation (STDEV) for each bias

The standard deviations of single values of precipitable water predicted by Model A (one-parameter model) extend to 5–7 mm in summer and autumn (about 1 mm less in the northernmost stations Sodankylä and Luleå). This is 30–32% of the station's summer mean and 42–48% of the autumn's mean. In spring the STDEV are 4–6 mm (44–49%) and in winter 3–5 mm (46–48%).

Using Model B, the STDEV are only 3.3–4.6 mm (18–22%) in summer, 2.6–4.0 mm (23–25%) in autumn, 2.4–3.4 mm (21-28%) in spring and 2.0–3.4 mm (31–34%) in winter.

Conclusions

We have examined the radiosonde midnight (00 UTC) low-resolution (15–30 levels) records of precipitable water *W* from a 17 station network in the Baltic region during the 14-year period 1989–2002. Our main findings are as follows.

1. In the Baltic region, single values of precipitable water never exceeded 45 mm and single monthly average values never exceeded 33 mm. The lowest single values were in the range of 1–2 mm.
2. The multiannual monthly means of *W* reach 25 mm in southern locations of the Baltic region, e.g. 25.9 mm in Legionowo in July and 25.1 mm in Wroclaw. In the northernmost stations (Sodankylä, Luleå) the lowest overall monthly means are 5.6 mm. The overall annual mean precipitable water in the region is highest in the southernmost station Wroclaw (16.2 mm); this value is considerably less than the global mean (25 mm).
3. The relatively small north–south extent of the Baltic area allows a representation of seasonal means of precipitable water as a linear function of latitude (Equation (2)). Attempts to consider precipitable water as a function of the geographic longitude degree, φ, and elevation above sea level, z, were not successful. Standard deviations of the one-parameter approximation (Equation (2)) are relatively high: 5–7 mm in summer (about 30% of the summer mean) and 3–5 mm in winter (almost 50%). However, due to its simplicity this approximation can be used in climatological studies, e.g. estimating absorption of solar and terrestrial radiation in water vapour for modelling the regional energy balance.

4. Availability of surface humidity data allows a much better approximation of precipitable water already using a linear function of surface water vapour pressure, e_0, and geographical latitude, φ (Equation (8)). Standard deviations of this two-parameter approximation are considerably smaller compared with the one-parameter approximation. The standard deviations are 3.3–4.6 mm in summer (18–22% of the summer mean) and 2.0–3.4 mm in winter (31–34%). In cases when more precise methods for the evaluation of precipitable water (radiosounding, solar photometry, GPS, microwave radiometry) are not available, this approximation can be recommended for optical studies of atmospheric aerosol particles. This approximation can also be used for compiling retrospective time series of precipitable water even for historical periods when radiosondes were not available.

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References

Bouma, H.R. and Stoew, B. (2001). GPS observations of daily variations in the atmospheric water vapor content. *Phys. Chem. Earth (A)*, **26**(6-8), 389–392.

Gueymard, C. (1998). Turbidity determination from broadband irradiance measurements: a detailed multicoefficient approach. *J. Appl. Meteorol.*, **37**, 414–435.

Güldner, J. and Spänkuch, D. (1999). Results of year-round remotely sensed integrated water vapor by ground-based microwave radiometry. *J. Appl. Meteorol.*, **38**, 981–989.

Ice Atlas (1982). *Climatological Ice Atlas*. Swedish Meteorological and Hydrological Institute, Finnish Institute of Marine Research, Norrköping.

Jakobson, E. (2004). Parameterization of atmospheric water vapor in the Baltic region. MSc Thesis, University of Tartu, See: http://www.utlib.ee/ekollekt/diss/mag/2004/b16663925/Jakobson.pdf.

Leckner, B. (1978). The spectral distribution of solar radiation at the Earth's surface – elements of a model. *Solar Energy*, **20**, 143–150.

Maurellis, A. and Tennyson, J. (2003). The climatic effects of water vapour. *Phys. World. May*, 29–33.

Mietus, M. (1998). The climate of the Baltic Sea basin. *WMO, Marine Meteorology and Related Oceanographic Activities, Report* no. 41. WMO/TD-No. 933.

Okulov, O., Ohvril, H. and Kivi, R. (2002). Atmospheric precipitable water in Estonia, 1990–2002. *Boreal Environ. Res.*, **7**, 291–300.

Peixoto, J.P. and Oort, A.H. (1992). *Physics of Climate*, American Institute of Physics, New York, p. 284.

Post, P., Truija, V. and Tuulik, J. (2002). Circulation weather types and their influence on temperature and precipitation in Estonia. *Boreal Environ. Res.*, **7**, 281–289.

Qiu, J. (2001). Broadband extiction method to determine atmospheric optical properties. *Tellus*, **53B**, 72–82.

Ross, R.J., Elliott, W.P. and Seidel, D.J. (2002). Lower-tropospheric humidity-temperature relationships in radiosonde observations and atmospheric general circulation models. *J. Hydrometeorol.*, **3**, 26–38.

Ruprecht, E., Schröder, S.S. and Ubl, S. (2002). On the relation between NAO and water vapor transport towards Europe. *Meteorologische Z.*, **11**(6), 395–401.

Sepp, M. and Jaagus, J. (2002). Frequency of circulation patterns and air temperature variations in Europe. *Boreal Environ. Res.*, **7**, 273–279.

Trenberth, K.E. and Smith, L. (2005). The mass of the atmosphere: a coinstraint on global analyses. *J. Climate*, **18**, 864–875.

WMO (1988). *Technical Regulations vol. 1 General Meteorological Standards and Recommended Practices* Appendix B, 1-Ap-B-3. World Meteorological Organisation, Switzerland.