Large-scale structures in middle atmospheric water isotopes $\text{H}_2^{16}\text{O}$ and HDO using Odin satellite data

Degree project in Meteorology 30Hp

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March 2012
Abstract

In this work Odin/SMR measurements of water vapour $\text{H}_2^{16}\text{O}$ (named $\text{H}_2\text{O}$ in the report) and its isotope $\text{HDO}$ for year 2002 - 2010 are used to study structures in the middle atmosphere indicating transport due to vertically propagating planetary waves. It is interesting to see if these structures show any interannual variability. A better knowledge of these processes can result in improved understanding of tracer transport due to the general circulation in middle atmosphere. The distribution of water vapour are connected to the middle atmosphere dynamics due to transport by the wind field. This together with that the life time of water is comparable to the time scale for dynamical transport processes in this part of the atmosphere makes water to a useful tracer for these studies.

Planetary Rossby waves are forced from land-sea contrasts or topography and are allowed to propagate vertically in the case of westerly zonal background winds. This correspond to winter hemispheric conditions and therefore northern hemispheric January and southern hemispheric July data are used. The northern hemisphere is also more favourable for planetary waves to develop due to higher topography and larger difference in land-sea distribution. The Rossby waves cause a displacement of the pressure surfaces and by that also the geopotential field which is related to pressure through the geostrophic wind relationship. Vertically propagating waves are therefore affecting horizontal and vertical transport processes.

A pronounced stationary wave one structure are found in the deviation from zonal mean $\text{H}_2\text{O}^\star$ in January around latitude 60°N with an amplitude of about 20% of the zonal mean values. Similar structures are also seen for $\text{HDO}^\star$ and in July data for both $\text{H}_2\text{O}^\star$ and $\text{HDO}^\star$ around 60°S, but these wave patterns are weaker and with lower amplitudes. When comparing the interannual variability in the wave one structures only $\text{H}_2\text{O}$ January data is used. This show both a phase-shift and a change in amplitude between the included years with maximum phase-shift of about 90°.

We are also comparing the structures in the deviation from zonal mean in water vapour $\text{H}_2\text{O}^\star$ with the deviation of geopotential height $Z^\star$. Because they are forced by the same mechanisms it could be possible to see a connection between them even though they might not be in phase. There is a shift in phase between $\text{H}_2\text{O}^\star$ and $Z^\star$ where the water vapour (except for one year) has a westward phase compared to the geopotential wave. But it is not possible to see any statistically significant correlation between them from this data.
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1. Introduction

Water vapour is the most important greenhouse gas and for that reason it is a very important constituent in the atmosphere. The fact that water in the troposphere can condense and form clouds further affects the radiative balance. Viewed from space the top of the clouds form a bright surface and an increased amount of clouds will then increase the Earth albedo. This will lead to that a larger part of the incoming solar radiation will be reflected back to space. A larger amount of clouds can also absorb more of the outgoing radiation from Earth surface, which then will be emitted both back to Earth surface and out to free space. How much of the radiation that is absorbed/emitted depends on the temperature i.e. on which altitude in the troposphere you find the clouds.

Water also plays an important part in several chemical reactions in the atmosphere e.g. ozone depletion. The primary source for ozone destruction comes from chlorofluorocarbons (CFCs), these long-lived species are stable in the troposphere but in the stratosphere they split up because of photodissociation due to a more energetic radiation and release chlorine. On the surface of polar stratospheric clouds (PSC) heterogeneous reactions take place where chlorine can go on and destroy ozone in catalytic cycles. An increased amount of carbon dioxide in the stratosphere due to climate change will lead to a decrease of temperature there. This together with an increased amount of water in the stratosphere (as a consequence of an increased amount of methane in the troposphere) will give better conditions for the formation of PSC and a larger amount of the species forming the PSCs will be available (Shindell et al., 1998). In the next step this will lead to a larger surfaces where these reactions can take place. The polar winter vortex act as barrier for transport from midlatitudes into the polar region, especially in the Antarctic where the polar vortex is more stable and the vortex air is well isolated. This barrier for transport of ozone into the polar region, together with reduced production of ozone due to the absence of sunlight during the polar night will also contribute to the creation of the observed spring time ozone hole.

When studying the middle atmosphere dynamics water vapour is very useful as tracer of dynamical transport because the chemical life time of water in this part of the atmosphere is comparable to the time scale for dynamical transport processes that take place here (Brasseur and Solomon, 2005). The amount of water vapour that is transported from the troposphere into the stratosphere is to a large extent controlled by the tropopause temperature and the flow of water vapour will follow an annual cycle. Most of the water vapour enters the stratosphere by ascent through the tropical tropopause transition layer (TTL). This region has very low temperatures and act as a "cold trap" (Holton et al., 1995) which strongly reduce the amount of water vapour due to freeze-drying and the ice particles that are formed in this process will then sediment leading to dehydration of water in the lower stratosphere. When the water is transported into the stratosphere it is marked with its entry mixing ratio, this is known as the atmospheric tape recorder effect (Mote et al., 1995), see section 2.2. The tropical region of the stratosphere is rather isolated because of the subtropical barrier preventing horizontal midlatitude mixing and leaving the rising air relatively unperturbed. Hence you can follow an air parcel by looking at its mixing ratio. The isotopic composition of the water is also important when you study the transport of water both in the troposphere and the stratosphere. In the troposphere the water will undergo several phase changes which leads to that the heavier isotopes HDO and H$_2^{18}$O will fall out through precipitation in a larger amount compared to the lighter H$_2$O. This will cause a depletion of heavy isotopes at higher altitudes in the troposphere. Methane is not depleted on heavier isotopes in the same rate as water. This will cause a difference in isotopic composition between the water that is transported into the stratosphere and the water that is formed in the stratosphere due to oxidation of methane. Even though there is a production of water from methane which increase the amount of water with altitude above the tropopause, the
stratosphere is very dry compared to the troposphere.

The general circulation driving the transport processes in the middle atmosphere are forced by the behaviour of planetary Rossby waves and gravity waves. These waves are propagating vertically in the atmosphere and due to the properties of the wave they break at different altitude levels. At the breaking momentum is deposited which affect the meridional circulation and wind direction in the atmosphere. This is part of the Brewer-Dobson circulation described in the theory section. By studying the large-scale structures in the middle atmosphere it is possible get a better understanding of the distribution of different chemical components, for instance ozone. It is difficult to observe these structures but one way of doing this is to use water vapour, the structures in the water distribution will reflect this circulation.

The strong connection between the middle atmospheric large-scale dynamics and the water distribution makes the water vapour to a useful tracer when studying these structures. Documentation of changes in the amount of water vapour and other important atmospheric gases can help us to understand how changes of this species affect the climate. The Odin satellite have now performed measurements of the atmosphere during the last 10 years and has become an important tool in these studies. In this work Odin measurements of the water isotopes \( \text{H}_2^{16}\text{O} \) and \( \text{HDO} \) over a time period from 2002 - 2010 will be used to study how the water are distributed in the middle atmosphere. It is interesting to see if it is possible to identify large-scale structures that can be related to the Brewer-Dobson circulation (see theory section) or planetary waves. Another interesting thing to examine is if these distribution pattern show any interannual variations.

2. Theory and background

2.1 General circulation in middle atmosphere

The region of the atmosphere that is usually termed the middle atmosphere consist of the stratosphere and the mesosphere. This part of the atmosphere differ from the troposphere in both temperature distribution and time scale for transport processes. In the troposphere temperature distribution is determined due to radiative heating of the surface with maximum surface temperature in the equatorial region and a decrease towards both the summer and the winter poles. When the warm surface air rise it is cooled adiabatic which results in a decrease of temperature with altitude. The large scale circulation act to compensate areas with low insolation by transport of heat from areas with high insolation.

Temperature in the lower stratosphere is to a large extent determined by the low temperature in upper troposphere. Above the tropopause the temperature in the stratosphere will increase with altitude due to absorption of ultraviolet radiation of ozone. This absorption give the stratosphere a positive laps rate with temperature maximum at the stratopause around 50 km. Above this level in the region termed the mesosphere the temperature will again decrease with altitude the lowest temperature throughout the whole depth of the atmosphere are observed at the mesopause around 80 km – 90 km (Wallace and Hobbs, 2006).

2.1.1 Stratosphere

Compared to the well mixed troposphere where temperature decrease with altitude, the stratosphere has a stable stratification which means that the temperature instead increase with altitude. This
stratification prevent vertical mixing in the stratosphere. Because of the low temperature of the tropical troposphere, the meridional temperature distribution in the lower stratosphere are characterized by a minimum at the equator and maximum at the summer pole and in midlatitudes in winter hemisphere. Above about 25 km (30 hPa) there is a uniform decrease in temperature from the summer to the winter pole (Holton, 2004). Figure 1 show temperature distribution in the middle atmosphere in June, northern hemispheric summer. Wind direction in the stratosphere is also changing between summer and winter hemisphere. This is a result from the temperature distribution above the 30 hPa level with temperature gradient pointing from the warmer summer pole towards the colder winter pole. The Coriolis force due to the rotation of the Earth will act to displace the flow to the right of the flow direction in the northern hemisphere (positive Coriolis parameter in NH) and to the left of the flow direction in the southern hemisphere (negative Coriolis parameter in SH). As a consequence of this there are easterly zonal winds in summer hemisphere and westerly zonal winds in the winter hemisphere showed in figure 2. The meridional temperature gradient also lead to that the zonal wind has a vertical gradient derived from the geostrophic wind and the thermal wind relationship, this is further explained in section 2.4. The result is an increase of the zonal wind with altitude.

![Figure 1. Zonal mean temperature (K) in middle atmosphere.](image1)

![Figure 2. Zonal mean wind (ms⁻¹) direction indicating easterly (E) wind in summer hemisphere and westerly (W) in winter hemisphere in the stratospheric region.](image2)
The change in wind direction between summer and winter hemisphere affect the general circulation in the middle atmosphere. Planetary Rossby waves forced from land-sea contrasts or topography in the troposphere are propagating horizontally with negative (easterly) phase speed. When these waves propagates vertically in the atmosphere they are limited to areas with westerly zonal background winds. This is because the waves can continue to propagate as long as its phase speed is lower than the background wind speed or if the phase speed is in the opposite direction compared to the zonal wind. When they reach a critical level where the phase speed equals the wind speed they will be irreversibly deformed and break. In the stratospheric summer hemisphere the wind direction is easterly and thus have the same direction as the Rossby phase speed. This will result in that the Rossby waves are not able to propagate vertically in this hemisphere. In the winter hemisphere where there instead are westerly winds the Rossby waves are allowed to propagate vertically. How far the waves are able to propagate are determined by the zonal mean state of the atmosphere and by the characteristic of the wave, its amplitude and zonal wave number. When the waves propagates vertically their amplitude will increase exponentially due to the decrease of air density with height. At some point there will be a limit where non-linear effects can not be neglected any more and the wave will become unstable and break (Holton 2004).

Waves that propagate vertically are travelling across isentropic surfaces and transfer angular momentum following the wave to a new altitude. When the wave break the angular momentum is deposited at that altitude. Deposition of momentum lead to a change in both relative and planetary angular momentum, this will cause a change in zonal wind and a meridional mass flux. Rossby waves carry westward angular momentum, at deposition this cause a negative drag that will slow down the westerly zonal wind and create a flow from the equator towards the winter pole (Shepherd, 2000). The region where this breaking occurs is called the “stratospheric surf zone” and is a part of the Brewer-Dobson circulation showed in figure 3. This figure give a picture of how the large scale circulation in the stratosphere works. The largest inflow of air from the troposphere to the stratosphere take place at the tropical tropopause. From here the air transports upward in the tropical stratosphere and then towards the pole where there is a downward motion. The sub-tropical jetstream creates a barrier which drives the circulation forced by the breaking of planetary waves to go around the jetstream instead of through it. In this way a circulation is created that transports the air across the midlatitudes and the descending motion occur at the pole.

Figure 3. Brewer-Dobson circulation showing the dynamics of troposphere-stratosphere exchange. Figure taken from Holton et al., 1995.
2.1.2 Mesosphere

The separation between the stratosphere and the mesosphere is defined by a change in vertical temperature gradient at the stratopause around 50 km altitude. Above this level the temperature will again decrease with altitude until the mesopause is reached around 85 km. In mesosphere the meridional temperature distribution deviates strongly from radiative equilibrium and the highest temperatures are found at the winter pole and lowest temperatures at the summer pole as indicated in figure 1.

This forcing arise from breaking of vertically propagating gravity waves which drive a meridional circulation. Compared to planetary waves, gravity waves can have both positive (westerly) and negative (easterly) phase speed and can thus propagate in both summer and winter hemisphere and reach all the way up into the mesosphere before they are deformed and break. The amplitude of the gravity waves are also smaller compared to the planetary waves. In winter hemisphere, where the stratospheric zonal wind is westerly, most of the gravity waves with positive phase speed will be filtered out when they reach an altitude level where their phase speed equals the speed of the background flow. The result is that mainly gravity waves with negative phase speed will propagate into the mesosphere where they break due to an increased amplitude caused by the decrease of air density. This breaking lead to deposition of momentum and a negative wave drag which lead to a weakening of the westerly background wind and a flow towards the pole. Above mesopause a sufficiently strong weakening of the zonal wind speed can also lead to a reversed wind direction in the mesosphere.

For summer hemisphere the stratospheric zonal wind is easterly and the conditions for propagation of gravity waves are reversed. Here will instead the gravity waves with negative phase speed be filtered out and the ones with positive phase speed have the possibility to reach into the mesosphere. Breaking of waves with positive phase speed will cause a positive drag and give a weakening of the easterly winds and due to the Coriolis force an equator ward flow. Together this result in a one-way circulation in the mesosphere from summer to winter pole. Due to mass conservation this imply up-welling with low temperatures as a result of adiabatic cooling of the rising air at the summer pole and down-welling at the winter pole (Shepherd, 2000).

Rossby waves are allowed to propagate trough the whole depth of the winter stratosphere, limited by the amplitude size for each wave. After the reverse in zonal wind direction in the mesosphere the remaining waves are filtered out depending on the local wind profile for different longitudes. This also apply to the breaking of gravity waves and may cause meridional differences which show up as wave structures in the zonal transport of atmospheric species in the mesosphere.

2.2 Water distribution in middle atmosphere

The amount of water vapour that is transported from the troposphere into the stratosphere is to a large extent controlled by the tropopause temperature. This is the lowest temperature that an air parcel will undergo on it is way up to the stratosphere. The very low temperature in the tropical tropopause layer (TTL) cause a large depletion of water vapour due to freeze-drying. The ice particles that are formed in this process will sediment which lead to the strong reduction of water. The mean value over a year that enters the stratosphere is only 3.7 ppmv (Kley et al., 2000) This will increase due to water production from oxidation of methane to about 7 ppm around the stratopause.
Because the tropopause temperature follows an annual cycle, this will give alternately higher and lower amounts of water vapour that enters the lower parts of the stratosphere. When the air then continue to rise, following the general circulation in the stratosphere, you can follow an air parcel by studying its mixing ratio. This is known as the tape recorder effect (Mote et al., 1995) and imply that when an air parcel pass the tropopause it is marked with its entry mixing ratio. When the air parcel then continue to rise it will have a different mixing ratio compared to the water that is produced in the stratosphere by oxidation of methane. This also makes it possible to calculate the ascent rate of the air (Mote et al., 1996).

Compared to the troposphere, the stratosphere is very dry. Beside the transport of water that takes place through the tropopause, there is a production of water in the stratosphere by oxidation of methane. Methane (CH\textsubscript{4}) is produced at the Earth's surface and is then transported into the stratosphere by the same way as water. In the troposphere it is a relatively long-lived species with life time greater than a year. In the stratosphere methane contribute to the water production by oxidation (1) and from reaction with OH (2).

\begin{align*}
CH_4 + 2O_2 & \rightarrow 2H_2^\text{16}O + CO_2 \\
CH_4 + OH & \rightarrow CH_3 + H_2^\text{16}O
\end{align*}

(1) \hspace{2cm} (2)

There is also a small part of water which comes from oxidation of H\textsubscript{2}, but except for the upper stratosphere this can be neglected.

The main sink process of water in the stratosphere is the reaction with O(^1D) (3) and for higher altitudes water vapour is destroyed by photodissociation (4), above 70 km this is caused by Lyman-\alpha radiation (\lambda=121.6 nm).

\begin{align*}
H_2^\text{16}O + O^1D & \rightarrow 2OH \\
H_2^\text{16}O + h\nu & \rightarrow H + OH
\end{align*}

(3) \hspace{2cm} (4)

Balance between the sources and sinks around the stratopause lead to an increased amount of water vapour in the stratosphere with maximum concentration around 50 km.

### 2.3 Isotopic composition

All atoms exist in different isotopic forms, but it is common to start out from a main isotope. From that one you define the different isotopes which can be either lighter or heavier than the main isotope depending on whether it contains more or less neutrons. The number of neutrons in a molecule does not change its charge but it changes its weight. For many important species the isotopic composition in the atmosphere differ from the composition at Earth surface. This makes it possible to trace the origin of an air parcel by looking at its isotopic composition and use that to study e.g. circulation and transport of different species. In general, heavier isotopes have lower vapour pressure. For water, the effect will be that when water evaporates from Earth’s surface, a larger amount of the main isotope will evaporate compared to the heavier ones. When the water instead condense, the heavier isotopes will condensate earlier than the lighter. This process will lead to a depletion of the heavier HDO and H\textsubscript{2}^\text{18}O higher up in troposphere compared to the conditions in the oceans.
The general source for tropospheric water vapour are the oceans, so the amount of water that evaporates from the oceans will decide the isotopic composition in the air above. To be able to compare different measurements with each other you start out from a standard value for the composition of oxygen and hydrogen, SMOW (standard mean ocean water). According to SMOW the relative amount of $\text{H}_2^{16}\text{O}$ in a sample of water is 99.73%, $\text{H}_2^{18}\text{O}$ 0.2% and HDO 0.03% (Rothman et al., 1987). To compare the isotopic composition of a sample, the ratio between the different measured mixing ratios for the different isotopes are determined.

$$\frac{(D/H)_{\text{sample}}}{2\kappa (H_2^{16}\text{O})} = \kappa (\text{HDO})$$

$kappa$=mixing ratio

For example the isotopic ratio for deuterium to hydrogen are calculated as

$$\delta D = 1000 \left( \frac{(D/H)_{\text{sample}} - (D/H)_{\text{ref}}}{(D/H)_{\text{ref}}} \right)$$

Standard values from SMOW for the reference ratio is $(D/H)_{\text{ref}} = 155.76 \times 10^{-6}$ and $(^{18}\text{O}/^{16}\text{O})_{\text{ref}} = 2005.2 \times 10^{-6}$ (Hageman et al., 1970).

The difference in vapour pressure also plays an important part when you study transport of water vapour from troposphere to stratosphere. Because the water goes through several phase changes in the troposphere, a larger amount of the heavier isotopes has fallen out through precipitation compared to the lighter ones, this will give a large $\delta D$-value. There will also be differences between different regions in the troposphere due to convection or evaporation from the surface.

For methane (CH$_4$) the depletion of heavy isotopes (CH$_3$D) in the troposphere is not so large as it is for water. The result is that when water is formed in the stratosphere by methane oxidation the amount of HDO compared to $\text{H}_2^{16}\text{O}$ will increase and the $\delta D$-value will come closer to the reference value from SMOW.

### 2.4 Geopotential height

The geopotential ($\Phi$) is defined as the work that has to be done against Earth's gravitational field to rise a mass of 1 kg from sea level to a certain altitude level in the atmosphere (Wallace & Hobbs., 2006)

$$\Phi(z) = \int_0^z g \, dz.$$  

Using the hydrostatic equation and the ideal gas law, $\Phi$ can be expressed in terms of pressure

$$\frac{\delta p}{\delta z} = -g \rho, \quad p = \rho RT$$

$$d \Phi = gdz = -\frac{RT}{p} dp = -RT \, d\ln p$$

$$\Rightarrow \Phi(z_f) - \Phi(z_0) = -\int_{z_0}^{z_f} RT \frac{dp}{p} = RT \ln \left( \frac{p_f}{p_0} \right)$$

the geopotential height ($Z$) defines the height above sea level and if pressure is used as vertical coordinate it defines the height for a certain pressure level. The geopotential are defined to be zero at sea level ($\Phi(z_0)$)
\[ Z = \frac{\Phi(z)}{g_0}. \]  

For synoptic-scale disturbances the equations of motion can be reduced due to scale analysis to an approximate balance between the Coriolis term and the pressure gradient term (Holton, 2004), giving the geostrophic relationship

\[ -fv \approx -\frac{1}{\rho} \frac{\partial p}{\partial x}, \quad -fu \approx -\frac{1}{\rho} \frac{\partial p}{\partial y}. \]

\[ f \equiv 2 \Omega \sin \phi \]

This expression include no time dependence and is thus a diagnostic expression which gives the relationship between the pressure field and the horizontal velocity field, but cannot be used to predict the evolution of the fields.

The geostrophic wind relationship \( (V_g) \) is derived by defining the horizontal velocity field as

\[ V_g = i u_g + j v_g, \]

\[ V_g \equiv k \times \frac{1}{\rho f} \nabla p \]

This horizontal wind relationship can instead be expressed using isobaric coordinate form (with pressure as vertical coordinate) by using relation (8) on the approximate horizontal momentum equation

\[ \frac{DV}{Dt} + f k \times V = -\frac{1}{\rho} \nabla p, \quad (8) \Rightarrow \frac{DV}{Dt} + f k \times V = -\nabla_p \Phi \]

where \( \nabla_p \) denote that pressure is held constant when using the gradient operator.

The geostrophic wind relation (11) can now be written in isobaric coordinate form with the horizontal wind components \( u_g \) and \( v_g \).

\[ f V_g = k \times \nabla_p \Phi \]

\[ u_g = -\frac{1}{f} \frac{\partial \Phi}{\partial y}, \]

\[ v_g = \frac{1}{f} \frac{\partial \Phi}{\partial x}. \]

The geostrophic wind define the horizontal wind field at a certain pressure level in the atmosphere, to find out how this wind change with altitude you can again look at relation (8), which show that the change in geopotential (\( \Phi \)) are depending on the temperature (\( T \)) in the layer between two pressure levels. This lead to that the presence of a temperature gradient affect the thickness of the layer between \( p_0 \) and \( p_1 \), if the temperature increase this imply that the the thickness also must increase due to the fact that the air expand when it becomes warmer. It also explain why the geostrophic wind must have a vertical shear in the presence of a horizontal temperature gradient. This can be visualized by the use of the thermal wind equation (14) which define the rate of change of the geostrophic wind with altitude.

\[ \frac{\partial V_g}{\partial \ln p} = -\frac{R}{f} k \times \nabla_p T \]
\[ u_T = -\frac{R}{f} \left( \frac{\partial \langle T \rangle}{\partial y} \right) \ln \left( \frac{p_0}{p_1} \right) \]
\[ v_T = \frac{R}{f} \left( \frac{\partial \langle T \rangle}{\partial x} \right) \ln \left( \frac{p_0}{p_1} \right) \]

\[ \langle T \rangle \text{= mean temperature of the layer} \]

From the zonal wind component \( u_T \) it can be seen that for an equatorward temperature gradient \( \partial T / \partial y \) is negative, because \( p_0 > p_1 \) the term \( \ln(p_0 / p_1) \) will be positive which imply that the zonal wind increase with altitude.

When planetary waves are forced from e.g. topography it will cause a displacement of pressure surfaces and the geopotential field, which is depending on pressure. Due to the connection with geopotential, vertically propagating waves will also affect the wind field. Atmospheric species, like water vapour will show distribution patterns due to advection of both horizontal and vertical winds. This should make it possible to observe related transport patterns in the distribution of water and geopotential even if they might not be in phase.

### 2.5 The Odin project

The Odin satellite is a cooperation between Sweden, Finland, Canada and France and was launched in February 2001 from Svobodny in Russia to an orbit at 620 km. The aim with the satellite was to make measurements for both astronomy and aeronomy. Odin was able to perform measurements of the atmosphere from space for altitudes up to 110 km, this will give new information about physical, chemical and dynamical processes in middle- and upper atmosphere. For instance ozone depletion and the impact of global warming on the middle atmosphere.

During the first years the time that it was possible to perform measurements was equally divided between the two research domains. In spring 2007 the astronomy studies was finished and after that the satellite is only used for studies of the Earth atmosphere. The life-time of Odin was guaranteed at the start to a minimum of 2 years but it is still working and the operation is now secured until the end of 2012. This long performance of Odin over more than 10 years makes the measurements important when studying the climate variability. The time corresponds to almost a solar cycle which extends for in average about 11.1 years.

The main instruments on Odin is a Sub-Millimetre Radiometer (SMR) which was built to be used for both astronomy and aeronomy observations and an optical spectrograph (OSIRIS) used for atmospheric studies. The SMR is using a 1.1 m telescope and works in the frequency bands between 486 GHz and 581 GHz and also in a band around 119 GHz (Urban et al, 2007).

#### 2.5.1 Observations of water vapour

Odin is placed in a Sun synchronous orbit, which mean that the satellite always pass the same point (relative to Earth) at the same point of time. Over time there has to be a small correction in this local passing time because the atmospheric drag causes a decrease of the orbit altitude over the years. In 2007 the satellite passed the equator about 45 min later compared to when it was launched. The orbit height is now around 600 km. The satellite are able to scan the limb of the atmosphere with different speed and perform observations for altitude levels between 10 km and 110 km. By its near polar orbit, Odin cover the latitude range from 82.5°S to 82.5°N the orbit period is 97 min which gives about 15 orbits per day. It can make 40 scans for each orbit and during the scans the
whole satellite is pointed towards the point of measurement. An observation period usually starts at 12 UT and last for 24h, but there is exceptions for some special measurements that require longer time interval.

The solar panels are always oriented within 35° of the sun, this guarantees that the energy production is at least 80% of maximum and that the telescope won't be exposed to direct sunlight (Murtagh et al, 2001). When the satellite follow its near polar orbit, for each orbit it will pass over both poles and there by scan all latitudes. Because the Earth is rotating under the satellite, after some time all longitudes will also be covered. The orbit does not reach all the way up to the pole so after a number of orbits both latitudes and longitudes are covered, but there will still be a hole right over both the north and south pole which is not covered by the measurements. To compensate for this the satellite is turned so that even the heights latitudes are covered. Because the telescope need to be kept in shadow from the sun, this is only possible to do in the winter hemisphere. The result is that when the scans are performed you get measurements all the way up to 90° latitude in winter hemisphere but only to about 82° latitude for summer hemisphere.

During the time that the satellite is used for observations of the atmosphere different species and areas are measured in fixed intervals. Measurements of the stratosphere are performed approximately every third day in an altitude range between 15 km and 70 km with focus on observations of ozone (O₃), chlorine monoxide (N₂O), nitrous oxide (N₂O) and nitric acid (HNO₃). For studies of middle atmosphere dynamic measurements of water isotopes are performed by vertical scans for altitudes between 7 km and 110 km. These measurements are made in weekly intervals. The same frequency bands are also used to measure stratosphere-mesospheric carbon monoxide (CO) and stratospheric ozone isotopes.

There are two emission lines for water one centred at 488.9 GHz and the other at 556.9 GHz. Lower and middle stratospheric studies are using the 488.9 GHz line, measurements of H₂¹⁶O and H₂¹⁸O are performed about one day per week at this frequency with a band width of 700 MHz centred at 488.9 GHz. At the same time a second radiometer is used in a band at 556.9 GHz, this emission line is very strong and can be used for measurements in the mesosphere. The 556.9 GHz provide information for altitudes from 40 km to 100 km. Every second orbit the 488.9 GHz is switched over to a band centred at 490.4 GHz to be able to do observations of HDO. These measurements gives information from about 20 km up to 70 km i.e. stratosphere and lower mesosphere. The emission lines for H₂¹⁶O and HDO are very close and are not as strong as the the other water line at 556.9 GHz. Because of this it is difficult to distinguish these lines from each other, but the high sensitivity of the SMR instrument make it possible to perform measurements for both of them (Urban et al, 2007). Together these observations provide a broad picture of the water vapour in the middle atmosphere. The upper altitude limit depend on the signal-to-noise ratio (SNR) i.e. the level of the signal power compared to the background noise and also of the decreasing volume mixing ratio of water.
3. Results

3.1 Data processing

The purpose with this work is to use Odin measurements of water vapour (H$_2^{16}$O), which will be termed just H$_2$O and its isotope deuterium (HDO) to look at the dynamics of the middle atmosphere. The interesting part is to see if there are any structures in the zonal distribution of water which can indicate wave activity in middle atmosphere and if there is any change in phase or amplitude of these waves between different years. The data that are used is the level-2 Odin/SMR data available from Chalmers university of technology\(^1\) for year 2002 until 2010 for both water vapour (H$_2$O) and the heavier isotope deuterium (HDO).

For this study of wave structures the winter time measurement is chosen because of the prevailing stratospheric winds that allow for vertical Rossby wave propagation. Rossby waves are forced from differences in land-sea distribution and from topography. The northern hemispheric winter is more favourable because there is a larger difference in land-sea distribution and higher topography than in the southern hemisphere. Thus we use northern hemispheric January and southern hemispheric July data both for the separate years and on the total for all nine years. Because the satellite was launched in February 2001 we chose to start from year 2002 so we get the same number of years for both January and July. Another reason to exclude 2001 is that the satellite instrument was being tuned during this first year which result in that the data is not very good. The near polar sun synchronous orbit of Odin cover all longitudes and latitudes between 82.5°S and 82.5°N.

Selection for down loading from the website is made using the time filter for the months in question for each year. These files contain data for all the species measured by Odin/SMR during the specified time period, band name IM_AC1c contain measurements of H$_2$O and IM_AC2c measurements of HDO. H$_2$O measurements retrieved from IM_AC1c are performed using the emission line at 557 GHz in frequency mode 19 which is the most optimized for measurements of the 557 GHz band and also the most used throughout the Odin mission. This band measures H$_2$O with good resolution in the altitude range 40 km – 100 km. For HDO measurements from IM_AC2c the frequency mode 17 is used with emission line 490.4 GHz, containing profiles in the altitude range 20 km – 70 km (Urban et al., 2007, Lossow et al., 2009). A summary is shown in table 1.

Figure 4 show the latitude-longitude distribution of H$_2$O in January using Odin/SMR data from 2002-2010. Even though the data is very noisy it is still possible to see the known water distribution with larger amounts of water in the summer hemisphere due to ascending of moist air and lower amounts due to descending of drier air in the winter hemisphere.

<table>
<thead>
<tr>
<th>Species</th>
<th>Frequency (GHz)</th>
<th>Frequency mode</th>
<th>Band name</th>
<th>Altitude range (km)</th>
<th>Altitude resolution (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>H$_2$O</td>
<td>557</td>
<td>19</td>
<td>IM_AC1c</td>
<td>40-100</td>
<td>3-3.5</td>
</tr>
<tr>
<td>HDO</td>
<td>490.4</td>
<td>17</td>
<td>IM_AC2c</td>
<td>20-70</td>
<td>3-4</td>
</tr>
</tbody>
</table>

**Table 1.** Summary of the included frequency bands from Odin/SMR.

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1. Odin/SMR website: http://odin.rss.chalmers.se
The data files for H$_2$O and HDO contain information about the position in latitude and longitude for the different profile scans and also at which altitudes the measurements are performed. Beside this there are information about the quality e.g. measurement response and error, also the time for the different measurements are specified. Only data with a measurement response larger than 70% are used. The measurement response tells how much information is taken from the a priori guess and how much is taken from the real measurement. A measurement response of 100% define that all information are from measurement and 0% that it is only a priori data from model calculations. The profiles for both H$_2$O and HDO contain measurements from ~ 7 km - 110 km, but there is a sharp limit where the sensitivity for the altitude range ends. Measurement response drops from > 0.7 down to ~ 0.1 for the next altitude, so there is no doubt about which part of the profiles that should be used and this restriction will give a good start for analysing the water profiles.

As can be seen in figure 4, the data appears very noisy. Some of the measured values are even unrealistic and may therefore be due to instrumental error or other disturbances. In order to get a better picture of the general wave activity, these outliers must be removed. To do so we use median absolute deviation (MAD). The median picks out the value that are in the middle of the measured data set. When using MAD, also the deviation from the median is calculated and then the median is again taken on the absolute deviation values:

$$\text{MAD} = \text{median}_i(\{|x_i - \text{median}(x)|\}) \quad (15)$$

Compared to the more common use of standard deviation (std), this method prevent that one single outlier will affect the data selection. Figure 5 show an example of the use of MAD compared to standard deviation. It can be seen that the median value (red line) are to the left of the mean value (green line). If taking the standard deviation this will displace the whole selection area to the right compared to the MAD. For a data set with a more pronounced outlier this difference between MAD and std will be larger and affect the data selection.
Figure 5. Comparison between MAD (red line) and std (green line) for a measured distribution with some pronounced outliers.

Table 2 shows the number of accepted profiles for each year before and after the outliers have been removed. Each profile is measured for a certain latitude-longitude combination, but for irregularly spaced altitudes in the range covered by the different species. The water vapour profiles are interpolated linearly in the vertical to fit our altitude scale, which range from 50 km to 90 km in steps of 2 km. The interpolated data are then placed in a homogeneous latitude-longitude grid covering 80°S to 80°N, 180W to 180E with a grid space of 10°×10°. We are also counting the number of scans that end up in each grid box. The median absolute deviation of the data is then calculated for each box, only the data that are within 5 times MAD are accepted and each box must include at least 2 profiles.

<table>
<thead>
<tr>
<th>Year</th>
<th>2002</th>
<th>2003</th>
<th>2004</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>2009</th>
<th>2010</th>
</tr>
</thead>
<tbody>
<tr>
<td>No. of scans before cuts</td>
<td>2732</td>
<td>1486</td>
<td>758</td>
<td>870</td>
<td>986</td>
<td>1231</td>
<td>2171</td>
<td>1953</td>
<td>2540</td>
</tr>
<tr>
<td>No. of scans after cuts</td>
<td>2643</td>
<td>1433</td>
<td>737</td>
<td>814</td>
<td>922</td>
<td>1144</td>
<td>2030</td>
<td>1842</td>
<td>2389</td>
</tr>
</tbody>
</table>

Table 2. Number of profiles of H₂O from Odin/SMR for year 2002-2010.

As could be seen in table 2 the number of accepted profiles with good data differ between the separate years. Even though year 2004 - 2006 contain about the same number of accepted profiles, it is only 2005 that has to be excluded when comparing the water distribution between the years. The reason is how the profiles are distributed over the latitude-longitude grid. Table 3 compare the distribution 2004 - 2006. From this it can be seen that many of the measurements in 2005 are concentrated in a few grid boxes, compared to 2004 and 2006 where the measurements are more equally spread over the whole grid.
### Table 3. Distribution of the measured profiles over the latitude-longitude grid for 2004 - 2006.

<table>
<thead>
<tr>
<th>Number of profiles/grid box</th>
<th>No. of grid boxes containing that No. of profiles after cuts</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2004</td>
</tr>
<tr>
<td>0</td>
<td>166</td>
</tr>
<tr>
<td>1</td>
<td>247</td>
</tr>
<tr>
<td>2</td>
<td>170</td>
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<td>0</td>
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<tr>
<td>16</td>
<td>0</td>
</tr>
<tr>
<td>17</td>
<td>0</td>
</tr>
</tbody>
</table>

Total number of grid boxes containing at least 2 data profiles: 216, 131, 308

3.2 The zonal mean state of middle atmospheric water

In the first analyse of how the water are distributed, the zonal mean for the measured concentrations of H$_2$O and HDO are plotted in figure 6. We want to see if we receive the known water distribution with low concentration in lower stratosphere, especially in the tropical region, and then an increase in concentration for higher altitudes up to ~ 50 km due to water production from methane oxidation. Above 50 km there will again be a decrease of water concentrations because of photodissociation. This is done to check that the data are treated correctly and the output is reliable and to be able to identify interesting areas that show distribution patterns due to wave activity.

Similar latitude-altitude cross-sections for H$_2$O and HDO using Odin/SMR data from 2002 in the altitude interval of 20 km – 70 km have previously been shown by e.g. (Zelinger et al., 2006) showing the clear structures with dehydration in lower stratosphere, especially in the tropical region and then an increase of concentration with altitude reaching a maximum around 50 km.
For a higher altitude range 70 km – 110 km, also using Odin/SMR data but from 2003 - 2008 (Lossow et al., 2009) showed the water distribution in the upper mesosphere and lower thermosphere during wintertime in both hemispheres. The mesospheric region between 70 km and 90 km show higher water concentration during summer due to the meridional circulation driven by the breaking of gravity waves resulting in up-welling of moist air over the summer pole and then down-welling of drier air over the winter pole. Above that level the circulation conditions is changed and the maximum in water concentration is found in the winter time polar region. This arise from a change in vertical wind direction from down-welling inside the polar vortex to up-welling above the vortex. When looking at the annual distribution in the polar region this increase in water concentration during winter is visible for altitudes above 90 km and with clear minimum around the equinoxes. This can be explained from that the photodissociation due to Lyman-α radiation, which is the major sink for water vapour above 70 km is more pronounced in summertime compared to the winter for both hemispheres.

In figure 6 the altitude stretches from 50 km - 90 km, so the maximum in concentration are expected to be around 50 km and then decrease with altitude. The higher concentration in summer polar region is due to the up-welling over the summer pole. Similar structures with larger amounts of water in summer polar region are seen in HDO during the same time period, but in this case only data up to 70 km are available.

![Zonal mean concentration for H2O in January from Odin/SMR data.](image)

3.3 Planetary wave structure in middle atmospheric water

To better see any structure indicating stationary waves in the water vapour we need to take the deviation from the zonal mean and compare for the defined altitudes.

\[
\text{H}_2\text{O}^* = \text{H}_2\text{O} - [\text{H}_2\text{O}]
\]

\[
\text{HDO}^* = \text{HDO} - [\text{HDO}]
\]

Where \([\text{H}_2\text{O}]\) and \([\text{HDO}]\) are the zonal mean. Measurements of \(\text{H}_2\text{O}\) and \(\text{HDO}\) from 2002 until 2010 in January and July are showed in figure 7 and 8. The upper panels show water concentration at latitude 60°N in January and for latitude 60°S in July, the lower panels show the corresponding deviation from the zonal mean.
As could be seen in figure 7 and 8 there is a clear structure in the distribution of H\textsubscript{2}O\textsuperscript{x} concentration around latitude 60°N in January, with generally lower concentrations in the area between longitude 180W - 0 and then over to generally higher concentrations from 0 – 180E. This distribution pattern is about the same for all of the included years when you look at one year at the time, but with slightly changed positions for the maximum and minimum in concentration. Figure 9 compare the measurements for year 2010 and 2008 both measured concentrations (H\textsubscript{2}O) and the zonal mean deviation (H\textsubscript{2}O\textsuperscript{x}). The distribution patterns are a bit noisy, but it is still possible to identify areas with minimum and maximum in concentration. This makes it natural to assume that it follows a stationary sine wave with wave number one. This also agree with other studies in the same subject, who also found a pronounced wave one pattern in H\textsubscript{2}O\textsuperscript{x} in the northern hemispheric winter (Gabriel et al., 2011).

A tendency to a similar pattern are seen in HDO\textsuperscript{x} in the total from 2002 - 2010 in altitudes between 50 km - 70 km but the structure is not as clear as for H\textsubscript{2}O\textsuperscript{x}, the concentration and the size of the amplitude of HDO deviation is also lower. This is also shown when looking at a single year in figure 9. Both 2008 and 2010 show clear structures in H\textsubscript{2}O\textsuperscript{x} but the distribution pattern are slightly changed between the years and especially in 2010 it is also possible to distinguish a wave-one structure in HDO but it is very noisy. Similar problems apply to July data for both H\textsubscript{2}O\textsuperscript{x} and HDO\textsuperscript{x}. One can see that there is a structure in the H\textsubscript{2}O\textsuperscript{x} distribution, but the variation is quite small and noisy. For H\textsubscript{2}O the structures is also displaced upward. The large difference between the result in January and July, especially for H\textsubscript{2}O, can be explained from the excitation of the planetary Rossby waves described earlier where the northern hemisphere is more favourable for the development of the planetary waves.
The maximum deviation in H$_2$O in January for the total of 2002 – 2010 is about 20% relative to the zonal mean and in July it is about 15%. Corresponding values for HDO is 15% in January and 10% in July. For the separate years the amplitude variation is larger and range between about 30-50% for January H$_2$O data and for HDO during the same time period the deviation amplitude is about 15-30%.

Figure 9. Comparison between the water distribution measured by Odin/SMR for year 2010 and 2008.

To be able to compare the distribution patterns showed in figure 7-9, we will use that a distribution depending on longitude can be described as a zonal mean flow with a Fourier series of sinusoidal perturbations (Holton 2004)

$$f(x) = \sum_{s=1}^{\infty} (A_s \sin k_s x + B_s \cos k_s x)$$  \hspace{1cm} (17)

These can be calculated by using a fast Fourier transform (matlab fft). The wave structures will then be given by using the amplitudes in the Fourier spectra and fit these ones together with the corresponding phase to a sine wave.

$$y(x,t) = A \sin (kx - \omega t + \phi)$$  \hspace{1cm} (18)

The amplitude (A) is defined as the deviation in concentration from the zonal mean and the phase-shift ($\phi$) as the longitude position of the minimum in concentration, $\omega$ is the angular frequency and k the wave number.

As was shown in figure 7-9 the water distribution seems to follow a wave one structure. Using an average of H$_2$O$^+$ for altitudes between 50 km and 80 km and latitudes from 50°N to 70°N along all longitudes in the fast Fourier transform (Matlab fft) an amplitude spectra is received. From this spectra for year 2002-2010 showed in figure 10, it is clear that one large amplitude is dominating over the rest of the spectra. This justifies the use of a simple sine wave to express the distribution of the zonal mean deviation in H$_2$O. The largest amplitude in the spectra should correspond to the amplitude of fitting the data to a simple sine wave. When comparing the wave-one structures only
H$_2$O measurements in January will be used. Since the structures are so weak for both HDO in January and the July data for both species and that the variations in concentration are so small, it will not be useful to fit a wave to these data. This is confirmed by studying the amplitude spectra for HDO in January where we do not receive the same structure with one clearly dominating amplitude. For HDO the largest amplitude in mixing ratio is $4.41 \times 10^{-11}$ and the 2nd amplitude is 55% of this size, compared to H$_2$O where the 2nd amplitude is just 27% of the size of the 1st amplitude. For the separate years the differences in HDO is even smaller and for some years it is not possible to take out one amplitude which is larger than the other. Further, because of the few accepted profiles in 2005 and their distribution over the grid this year do not contain enough data to give a qualified result.

Beside the amplitude spectra, figure 10 shows the measured concentration H$_2$O together with the fitted sine wave. It can be seen that the largest amplitude from the Fourier transform agree with the sine wave amplitude of 0.37 ppmv, the phase of the wave is -74.67º. Both the averaged H$_2$O concentration and the fitted wave give a good representation of the total concentration showed in figure 7.

![Figure 10](image-url)  
**Figure 10.** Amplitude spectra from the Fourier transform (fft) and the fitted sine wave for year 2002-2010.

We also compare the stationary waves in H$_2$O with reanalysis of geopotential height deviation from zonal mean

$$Z^* = Z - \bar{Z}$$  

Geopotential height data (m) are received from ECMWF, ERA-Interim (Dee et al., 2011) for January 2002 - 2010 at 1 hPa (~45 km). The data was averaged over latitude 50ºN-70ºN and the wave component was determined. The 1 hPa level is the highest altitude with reliable data for ERA-Interim. In geopotential height there is a phase-shift with altitude, this is the reason why it is not possible to take a vertical average here. Gabriel et al., 2011, showed a similar phase-shift in water vapour distribution as well up to 50 km, but above that level the wave are more steady in position. This allows the vertical averaging in water vapour.

In all other aspects phase and amplitude for Z are calculated in the same way as for the water concentration. As was mentioned in the theory part, in should be possible to observe related structures in the deviation from zonal mean, since the wind fields associated with the geopotential height cause the wave in water vapour. Figure 11 show H$_2$O and Z in 2003, in the break level at 1 hPa / 45 km between the figures the concentration pattern seem to follow quite well. Also the phase-shift with altitude for geopotential height is visible and the more steady phase in water above 50 km.
Figure 11. Measured concentration $\text{H}_2\text{O}^+$ and $Z'$. Note that upper altitude level 1 hPa in geopotential height corresponds approximately to the lower level 50 km in water concentration.

3.4 Interannual variability in mesospheric wave-one

As could be seen in figure 7-9 the structure in $\text{H}_2\text{O}^+$ looks slightly different between the individual years and also compared to the over all structure for 2002 - 2010. The calculated sine waves will be used to compare the interannual variations of the wave one structures in water vapour and geopotential height. Figure 12 shows the measured concentration $\text{H}_2\text{O}^+$ together with the fitted sine wave for year 2008 and 2010. The differences in the distribution patterns that was seen in figure 9 are now visualized by the sine wave with a phase-shift from -17.95 (2008) to -111.77 (2010), see table 4, and the amplitude decrease from 0.52 ppm in 2008 to 0.36 ppm in 2010. From the specified values in table 4 it can be seen that the difference between 2008 and 2010 also represents the maximum range for the phase-shift. This give a maximum phase-shift of 93.82 degrees over the included nine years.
Figure 12. Measured concentration $\text{H}_2\text{O}^+$ for 2008 and 2010 together with the fitted sine wave.

Figure 13. Phase and amplitude in $\text{H}_2\text{O}^+$ and $Z^*$ for the fitted sin wave for year 2002-2010.

In figure 13 phase-shift and amplitude for the fitted sine wave are plotted for each year from 2002 until 2010 for both water vapour and geopotential height, the missing data for $\text{H}_2\text{O}^+$ in 2005 is because of the few accepted profiles that year. Largest amplitude at 0.56 ppmv are found in year 2002 and the smallest at 0.34 ppmv in 2009. In figure 13 we can also notice that the increase and decrease in phase and amplitude for $\text{H}_2\text{O}^+$ follow each other, which they necessarily don't have to do. The same tendency are also seen in $Z^*$, one explanation could be that certain phases are more favourable for large amplitudes to develop then others.

A comparison between the phase-shift in water concentration and geopotential height is shown in figure 14 and the values are then specified in table 4.
At a first look the phase in H$_2$O$^x$ and Z$^x$ in figure 14 seem to be quite uncorrelated. But if we look closer to the values for each year and the difference between them, the maximum difference is 86° out of 360°, i.e. a phase-shift of about 90 degrees. For all years, except 2004 the H$_2$O$^x$-wave have a phase located to the west of the Z$^x$-wave.

The received correlation coefficient is -0.07 i.e. there is no correlation between the phase in H$_2$O$^x$ and Z$^x$, the minus sign denote anti-correlation. This correlation coefficient is not statistically significant and it is therefore not possible to say if there is a connection. The fact that the wave pattern is fairly stationary also make it more difficult to detect a clear relationship between H$_2$O$^x$ and Z$^x$.

The connection between the wind field and geopotential described in the theory section can be part of the explanation for the phase difference between H$_2$O$^x$ and Z$^x$. If the geopotential are described as a sine function

$$\phi = A \sin(x)$$  \hspace{1cm} (20)

the geostrophic wind component are defined as

$$v_g = \frac{1}{f} \frac{\partial \phi}{\partial x} = \frac{1}{f} A \cos(x)$$  \hspace{1cm} (21)

If only this transport term are considered it will result in a phase-shift of 90° between the geopotential field and the water vapour, which is transported by the wind field. The observed westward phase-shift agree with Gabriel et al., 2011, who showed a similar phase difference with water vapour about 90° to the west compared to the geopotential height. In the same work it was shown that the advection by geostrophically balanced winds are the primary contribution to the
wave structures observed in H$_2$O$^\cdot$

Another aspect that may affect the result when deriving the phase values is the shift in altitude between H$_2$O$^\cdot$ and Z$^\cdot$ data. Also the quality of the ERA-Interim data for this altitude can be discussed. It is know that this reanalysis have good quality for e.g. the 500 hPa level and that the quality decrease for higher altitudes. The 1 hPa level is the highest altitude that should have a reliable quality. The conclusion is that we do not see a statistically significant correlation between water vapour and geopotential height from this comparison, even though their structures should be forced by the same mechanisms.

4. Discussion and Conclusions

Satellite data from Odin/SMR measurement of water vapour H$_2$H$^{16}$O (named H$_2$O) and the heavier isotope HDO for year 2002 – 2010 have been used to examine structures indicating transport due to planetary waves in the middle atmosphere. A clear wave one structure have been seen around latitude 60°N in January and 60°S in July in the altitude range of 50 km - 80 km, showed in figure 7-9. This is the region of the upper stratosphere and the mesosphere. It was also showed that the wave structures was more pronounced in the northern hemispheric winter compared to the southern hemisphere, this is a consequence of stronger forcing of Rossby waves in the northern hemisphere due to larger land-sea contrasts.

The received level-2 Odin/SMR data was quite noisy and it required several steps to take out good data that could be used to show the water distribution. Two things that was used was the measurement response limit of at least 70% to make sure that profile values really came from the actual satellite measurement and not from included model calculations. The other one was the Absolute median deviation (MAD) which prevent that outliers will affect the data.

When comparing the distribution between the included years a picture is formed of a stationary wave with wave number one with a minimum in water concentration for the west longitudes and maximum in the east longitudes. The deviation in H$_2$O January data showed an amplitude of about 20% compared to the zonal mean and in HDO the amplitude was about 15%. For July the structure is weaker and the amplitudes are about 15% in H$_2$O and 10% in HDO. Because of this only January data for H$_2$O is used when the interannual variability are examined. This variability showed a shift in both phase and amplitude for the sine wave fitted to the data. The total range for the phase-shift was about 90° and the amplitude variation was between 0.56 ppmv and 0.34 ppmv.

Almost the same structures is visible in the comparison with the deviation from zonal mean in geopotential height (Z$^\cdot$). Except for one year, the wave structure in H$_2$O$^\cdot$ have a westward phase-shift compared to Z$^\cdot$ with a maximum shift of 86°. This agree with the connection between geopotential height and the wind field which transports water vapour described in the theory section, but it was not possible to find a statistically significant correlation between them.

In figure 7-9 it can be seen that the HDO structures are still quite noisy, there is a tendency to similar structures as in H$_2$O but these are quite weak. The lower amount of HDO in the atmosphere compared to H$_2$O can result in that the HDO data are covered by the background noise, especially the zonal mean deviations. An increase in the number of measurement by using a longer time period of e.g. three month instead of one can perhaps reduce the signal-to-noise ratio and it might be possible to see a pronounced wave structure in HDO as well.
There is quite large differences in the number of accepted profiles for the included years, specified in table 2. One way to deal with this problem is to use the same method as was described for improvement of the HDO signal, i.e. increase the data set by using the whole winter season of three month, DJF in northern hemisphere and JJA in southern hemisphere. The Odin satellite follows a sun-synchronous orbit performing 15 orbits/day, this mean that during one day of measurements it will cover every 24th longitude. The longitude coverage is slightly displaced between the days so after some time it will cover all longitudes with profiles. The time that it is possible to perform measurements are divided between the different species of interest and the number of days with water measurements can differ between the months. So a longer time period will make sure the a larger area of the Earth are covered. There are also some special campaigns of measurements during longer time periods that changes the planed schedule and these can reduce the number of measurements for other species during that period, which was the case in January 2005. Another way to increase the number of measured profiles is to include an other water frequency mode. The mode 13 also measure H$_2$O at 557 GHz that can be used for this purpose. The fact that there are less measurements performed of HDO than H$_2$O during one month makes the HDO data more sensitive when using a shorter time period.

**Acknowledgements**

I would like to thank my supervisors, Heiner Körnich, Linda Megner and Kristoffer Hultgren for great support during this work. A special thank to Linda who has taken her time helping me to struggle through the difficulties with the data selections. I would also like to thank Stefan Lossow for helping us with which data files that contained useful water measurements to start out from and for great discussion about the Odin/SMR measurements.
References


