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The impact of thin water vapor layers on CHAMP radio occultation measurements

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Abstract.
CHAMP radio occultation data is analyzed to determine causes for signal tracking loss at altitudes above the Planetary Boundary Layer (PBL). This frequently happens within the Intertropical and South Pacific Convergence Zones. We focus on events with a sudden Full Spectrum Inversion (FSI) amplitude drop over a narrow atmospheric layer, as for example caused by strong atmospheric refractivity gradients. Events show a clear distinction between land and sea based observations. At high latitudes and mountainous regions, sudden FSI amplitude drops are caused by GPS satellite settings. Low (mid) latitudes sea based events also show maxima around 7 km (5 km) and 2.5 km. Events near 2.5 km are found at the upper range of the PBL, possibly involving ducting. Wave optics simulations applying CHAMP receiver characteristics and simulated refractivity profiles show that such events can readily be caused by strong refractivity gradients below the ducting conditions. Hence, around 7 km, events are caused by the relatively stable temperature lapse rate leading to enhanced water vapor layers and radiative cooling, associated with cumulus congestus clouds. The required signal carrier-to-noise density ratio sufficient to avoid tracking losses is 55 dB Hz for low latitude observations. In the lower atmosphere, near the PBL top, 60 dB Hz is required in the presence of strong vertical gradients in refractivity. CHAMP operates with about 50 dB Hz at low latitudes.

1. Introduction

Radio occultation instruments using GPS signals allow to measure the changing vertical refractive index of the atmosphere \cite{Kursinski et al., 1997; Rocken et al., 1997}. Further processing of these measurements allows to derive temperature and water vapor profiles, as well as surface pressure. Several radio occultation instruments have already shown the potential of this measurement technique. For an overview on past and future missions please refer to \textit{Yunck et al.} [2000].

The GPS signals are observed in limb sounding geometry, hence the measurement requires the refractivity profile to be sufficiently smooth so that the signal observed is not bent too strongly and tracking maybe lost. Strong refractivity gradients are usually associated with changes in the water vapor concentration. In fact, the amplitude variations of the Full Spectrum Inversion (FSI) processing \cite{Wickert et al., 2005} clearly shows the global water vapor climatology even at high altitudes.

Figure 1 (left) shows the normalized amplitude variations of the CHAMP instrument measurement around 7 km for 4 Degree latitude longitude boxes. The FSI technique is a recent development to process radio occultation data and resolves the influence of multi path on the signal \cite{Jensen et al., 2003}. The FSI amplitude is a measure of the signal strength received; for a spherically symmetric atmosphere it depends on the refractivity gradient and the receiver noise (for an example of the FSI amplitude, please refer to \textit{Jensen et al.} [2003]). The FSI amplitude itself is very variable and the mean over a 1 km wide vertical interval centered at 7 km was used to calculate the mean and standard deviation. Variations are then calculated as the averaged standard deviation in a box over the mean amplitude in that box.

When GPS based radio occultation measurements were first proposed it was hoped that the tracking would regularly penetrate the PBL. The right hand side of Figure 1 shows the

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number of observations in a latitude longitude box where processing was terminated above 3 km, thus well above the Planetary Boundary Layer (PBL), which is usually found below about 2.0 km. For CHAMP this altitude is determined by the smoothed FSI amplitude. If this drops below 0.5, the signal-to-noise ratio is considered to be too low for further processing. Figure 1 (right) shows that this target has not yet been met for the current radio occultation instrument CHAMP. It clearly shows that a lot of measurements are lost in the Intertropical Convergence and South Pacific Convergence Zones (ITCZ and SPCZ).

One reason for the signal tracking loss are strong refractivity gradients. In fact, one limit to signal tracking is reached when so-called ducting (or super refraction) occurs in the atmosphere. Ducting occurs when the refractivity gradient $dN/dr$ with respect to altitude $r$ is less than about $-160 \text{ km}^{-1}$. Ducting in radio occultation data has been discussed by Sokolovskiy [2003], in the atmosphere it is mostly located at the top of the PBL or, for polar regions, near the surface [Patterson, 1982; von Engeln and Teixeira, 2004]. Nevertheless, radio occultation measurements performed by the CHAMP satellite are terminating often above the PBL and even at altitudes higher than 5 km. In total, almost 41 % of occultations at low latitudes are terminating above 3 km. These terminations cannot be explained by ducting conditions at these altitudes, but are caused by strong refractivity gradients below the ducting criteria.

The focus of this paper is thus twofold: on the one hand we investigate why is it that a substantial amount of observations are not reaching the PBL and what instrument modifications are necessary to assure tracking into the PBL. On the other hand we use some characteristics of the FSI amplitude to study thin layers of water vapor in the atmosphere. The sensitivity of radio occultation data to thin atmospheric layers of water vapor has already been demonstrated in de la Torre Juárez and Nilsson [2003], although by directly using the refractivity profile determined from the radio occultation measurement and a simple model of atmospheric refractivity. The method proposed over here is based on the actual processed GPS satellite signal using the FSI amplitude, thus is more robust when strong vertical gradients in refractivity are present, which can lead to biases in the refractivity profile [Ao et al., 2003].

Our article is thus structured as follows: Section 2 introduces the conditions under which strong refractivity gradients can develop and also analysis radiosonde data; Section 3 looks into simulations of radio occultation observations under different refractivity gradients; Section 4 discusses the data analysis technique, and Section 5 closes with the conclusion.

2. Atmospheric Refractivity Gradients

Refractivity $N$ is calculated at each atmospheric level following the formula given by Smith and Weintraub [1953]:

$$N = 77.6 \frac{p}{T} + 3.73 \times 10^5 \frac{e}{T^2}$$

(1)

where $p$ is the atmospheric pressure [hPa] at that level, $T$ the atmospheric temperature [K], and $e$ the water vapor partial pressure [hPa]. The influence of the ionosphere is ignored.
Atmospheric refractivity gradients have been analyzed theoretically, where the impact of temperature, water vapor, and pressure has been assessed. Additionally, radiosondes were analyzed for the occurrence of water vapor gradients.

2.1. Theoretical Calculations

Refractivity gradients are mainly caused by the varying temperature and water vapor profile and to a lesser extent by the pressure. Performing the derivative $dN/dr = N'$ of Eq. 1 results in 4 contributing terms:

$$N' = 77.6 \cdot p' \cdot \frac{1}{T}$$

$$-77.6 \cdot T' \cdot \frac{p}{T^2}$$

$$+ 3.73 \times 10^5 \cdot e' \cdot \frac{1}{T^2}$$

$$-3.73 \times 10^5 \cdot T' \cdot \frac{e}{T^3}$$

Figure 2 shows examples of the 4 terms of Eq. 2, for very strong refractivity gradients in ERA 40 and ERA 40 like ECMWF re-analysis fields [Simmons and Gibson, 2000]. Gradients of both profiles are reaching ducting conditions of $N' \leq -160 \text{km}^{-1}$. The four terms on the right hand side of equation 2 are labeled $p', T', e', T'_2$.

The first term reflects the hydrostatic equation. It slightly decreases with decreasing altitude, having a mean value of about $-30 \text{km}^{-1}$. The dominant contribution for a mainly temperature based refractivity gradient comes from the second term, caused by a surface based temperature inversion [von Engeln and Teixeira, 2004]. This term generally increases $N'$ slightly when no temperature inversion is present, since temperature decreases with altitude. A lowering of $N'$ due to this term only happens at the altitude of an inversion in temperature, even for a mainly humidity based strong refractivity gradient at the top of the PBL where temperature is reduced. The third term represents the major contribution for a mainly humidity based gradient, here (Figure 2, right) caused by the strong decrease in water vapor pressure at the top of the PBL. This also helps to understand the correlations between the FSI amplitude and the global water vapor distribution, as found in Figure 1; temperature based gradients develop under dry conditions and this third term is negligible. The fourth term is similar to the second one but is weighted with $1/T^3$ and $e$ instead of $p$, it is thus smaller.

Since temperature inversions are generally found near the surface [von Engeln and Teixeira, 2004], they cannot explain the signal tracking loss at higher altitudes. Strong refractivity gradients will mainly develop based on humidity gradients, although temperature gradients might contribute, in particular caused by radiative cooling near the top when clouds are present. Figure 3 shows the required theoretical increase in relative humidity (RH) at a specific altitude to cause strong refractivity gradients for 2 cases and a tropical standard atmosphere [Anderson et al., 1986].

We assume a layer thicknesses of 50 m over which the change happens linearly. The RH value is increased from the profile background at the bottom of this layer until it hits the given refractivity gradient or saturation. The RH value at the top remains at the background level. The required increase in RH at a particular altitude to reach the gradient is given in the figure. The layer is then moved through all altitudes as given in the figure. Also shown are the effects of different associated linear temperature changes over the layer. A 50 m thick layer was taken since analyzed low latitude AWI radiosonde data [König-Langlo and Marx, 1997] shows strong refractivity gradients over vertical layers of about 50 m at altitudes above 3 km. RH was calculated over water for temperatures above 273 K and over ice below. The increase is terminated for RH reaching 100 %. The derivation is based on a 3-point Lagrangian interpolation. The asymptotic behavior visible at the upper altitude is caused by the decreasing impact of water vapor on refractivity. Note that small scale variations are caused by numerical noise in the derivation since with each variation in RH and temperature, the hydrostatic pressure profile must also be recalculated.

Figure 3 shows that the refractivity gradient depends on four factors: 1. the altitude of the layer; 2. the layer thickness over which the change happens; 3. the associated RH change; 4. the associated temperature change. In the lowest kilometer, strong refractivity gradients can readily be generated by a small RH increase; to generate a gradient of $-100 \text{km}^{-1}$ here, a RH increase of only 5 % is sufficient, in order to reach ducting conditions a 10 % increase is required or a sudden temperature decrease (Figure 3: bottom). A temperature increase on the other hand requires a larger increase in RH to cause a strong refractivity gradient. This figure also shows that ducting above about 4 km will always require a RH increase over a sharply defined layer and that ducting can theoretically be found up to the mid-troposphere, as already estimated by Kursinski et al. [2000]. This altitude moves further up for a gradient of $-100 \text{km}^{-1}$ by about 1 km.

Figure 3 is based on theoretical computations of RH and/or temperature variations over a sharply defined layer. Within the lowest few kilometers, vertical mixing in the PBL over the ocean is very strong and thus prevents the formation of these layers with a sharp gradient. These gradients of water vapor on the other hand are a regular feature of the free troposphere [Newell et al., 1999] or at the top of the PBL [Duynkerke et al., 1999; Stevens et al., 2001; Siebesma et al., 2003], where a strong decrease in water vapor exists.
Figure 2. Individual contribution of the 4 terms on the right hand side of Eq. 2 to the refractivity gradient, Left: for a mainly temperature based gradient (time: 15/07/2002 00 UT, location: 0.0 W, 10.5 S), Right: for a mainly humidity based gradient (time: 15/07/2002, 18 UT, location: 1.5 E, 81.0 S). Note: different altitude ranges.

Figure 3. Temperature and relative humidity change required for a tropical atmosphere and a 50 m thick layer located at that altitude to generate a refractivity gradient of $-100 \text{ km}^{-1}$ (left) and of $-160 \text{ km}^{-1}$ (ducting, right).
2.2. Radiosonde Investigation

At higher altitudes strong refractivity gradients are mainly driven by changes in RH as shown above. We use AWI (Alfred-Wegener-Institute, Bremerhaven, Germany) radiosondes to search for such strong RH changes. The dataset starts in 1982 and ends early 2004. In total, there are more than 6,500 measurements available, with about 4,100 at high latitudes, generally above the Atlantic ocean. The vertical resolution of this dataset is around 25 m to 50 m. A general consistency check of the RH data was performed first, profiles with variations of more than 50% in RH found between two datapoints were removed; these points were generally found above 10 km (indicating problems with the RH sensor) or at polar latitudes (thus not relevant for this study).

Figure 4 shows the frequency of occurrence of strong gradients of RH over a 50 m thick layer. Latitude bands are binned in 30°. For each RH band, only the found maximum value in a radiosonde profile enters the statistic. A few radiosondes with decreases of 40% to 50% are still found at altitudes around 6 km in the tropics and changes of 30% to 40% occur about 1% of the times. Mid and high latitude profiles also show strong RH gradients, although the generally lower temperatures (and thus water vapor concentration) will limit the impact on the refractivity. There is also a minimum visible in occurrence of the 30% to 50% changes for altitudes around 3.5 km at low latitudes. Temperature decreases (not shown) over a 50 m thick layer are usually below 0.5 K. About 4% of AWI radiosonde profiles show decreases of 0.5 K to 1.0 K for altitudes above about 1.5 km, independent of the latitude band. Sub-percentage of profiles fall into the 1.0 K to 1.5 K band at these altitudes. Below 1.5 km, decreases of up to 3 K over a 50 m thick layer can be found. The lapse rate for a typical tropical profile is about 0.3 K over a 50 m layer.

Such sudden decreases in water vapor have been the focus of several articles in the context of atmospheric layers, e.g. Newell et al. [1999]; Thouret et al. [2000] who analyze measurements of commercial aircrafts to characterize the fine laminar structure in the tropopause. Even earlier, other authors analyzed soundings within the TOGA COARE field program for dry tongues [Mapes and Zuidema, 1996]. A layer with an enhanced level of water vapor also shows radiative cooling. Thus temperature and water vapor terms are contributing to the refractivity gradient, as shown in Figure 2 (right). Analyzed data shows only very little seasonal behavior in the aircraft data; more layers are found at northern mid-latitudes in summer than in winter, and a maximum of layers per profile in spring for tropical Asia. Layers are mostly found around 5 km to 7 km in the tropics [Newell et al., 1999], an altitude where cloud tops are frequently observed [Zuidema, 1998; Johnson et al., 1999].

Mapes and Zuidema [1996] argue that dry tongues can inhibit the penetration of would-be penetrating convective clouds, thus preventing deep convection and creating preferential altitudes for cloud formation. Brown and Zhang [1997] also reported on an extremely dry mid-troposphere with few deep clouds. Johnson et al. [1999] argues that there are 3 prominent cloud types in the tropics: shallow cumulus, cumulus congestus, and cumulonimbus, where the top of cumulus congestus are found near the 0°C level. Earlier articles show that temperature and moisture perturbations are a common feature near this level [Mapes and Zuidema, 1996; Johnson et al., 1996], attributed to a relatively stable temperature lapse rate. Yasunaga et al. [2003] relates these enhanced cloud layers to cumulus convection that transports boundary layer air directly to the 350 hPa to 600 hPa level (about 5 km to 8 km), especially in regions of warm tropical sea surface temperatures in the ITCZ and the SPCZ. Altitudes inbetween these relatively stable temperature lapse rate layers are less likely to show cloud formation and are usually found between about 600 hPa to 750 hPa (about 2.5 km to 5 km) [Zuidema, 1998; Yasunaga et al., 2003].

3. Simulations of Radio Occultation Observations

As mentioned above, layers with strong refractivity gradients in AWI radiosonde data at higher altitudes are always rather thin; they are caused by a sharp gradient in the water vapor concentration at the top of layers with enhanced water vapor or at the bottom of dry tongues. These water vapor layers are several hundred meters thick [Newell et al., 1999], but the refractivity gradient is only affected by the transition from a low background to an enhanced level. To simulate these transitions and assess their impact on radio occultation FSI amplitude and refractivity biases, we generate smooth refractivity profiles based on the formulas presented in Sokolovskiy [2001a]:

\[
N_{A}(r) = 400.0 \times \exp\left(-\frac{r}{8\text{ km}}\right) \\
N_{b}(r) = N_{A}(r) \times \left[\frac{1 - 0.05 \times \frac{x}{\pi} \tan^{-1}\left(\frac{r - 6\text{ km}}{0.05\text{ km}}\right)}{1}ight]
\]

where \(N_{A}\) is a smooth refractivity profile over altitude \(r\), and \(N_{b}\) is the modified refractivity profile with a perturbation at 6 km. The factor \(x\) was set to 2 in the original publication but has been modified here to vary between values of 0.0 and 8.0 to generate a range of refractivity gradients at 6 km. The generation of refractivity profiles using Eq. 3 has several
Figure 4. Frequency of occurrence of sharp RH gradients in radiosonde data at a particular altitude for different gradients over a 50 m vertical layer, separated by latitude bands. Vertical bins are 1 km, total number of entering radiosondes is given in upper right corner. The legend gives the different RH gradients intervals used for sorting.
advantages over the variation of temperature and water vapor values at a certain altitude (as performed in Section 2), e.g. it requires no regeneration of a hydrostatic profile, is less noise sensitive, and results directly in smooth refractivity profiles.

The radio occultation simulator used in this study was developed at the GeoForschungsZentrum Potsdam, Germany, where CHAMP radio occultation observations are operationally processed to generate profiles of refractivity, temperature, and water vapor. It has been used in a number of publications, e.g. [Beyerle et al., 2003, 2004, 2005; von Engeln et al., 2005]. We use estimates of CHAMP like receiver and error settings, with signal carrier-to-noise density ratio (also called $CN_0$) of 50 dB Hz, typical for low latitude observation. An analysis of CHAMP low latitude observations showed signal carrier-to-noise density ratio varying between about 46 dB Hz to 52 dB Hz with a mean at about 50 dB Hz.

The signal carrier-to-noise density ratio $SCN$ [dB Hz] is calculated from the CHAMP signal-to-noise ratio $SNR$ [1] values via:

$$SNR = \sqrt{2 \times T \times 10^{SCN/10}}$$

with the integration time $T$ (which is generally set to 1 s to derive SNRs for the Black Jack receiver) [Montenbruck and Kneis, 2003].

The atmosphere is assumed to be spherically symmetric. CHAMP operational processing is terminated when the normalized, smoothed FSI amplitude drops below 0.5. This setting has not been applied here to analyze tracking to lower altitudes, thus we use the full amplitude and investigate the reduction in FSI amplitude depending on different refractivity gradients. It has been found in simplified calculations that strong gradients can significantly reduce the FSI amplitude. The amplitude often does not recover from this strong reduction below the strong gradient and stays below a normalized value of 0.1, a fact found in simulations as well as in CHAMP observations.

The detection of sudden signal loss or strong reduction was performed by first smoothing the FSI amplitude data over a 500 m vertical interval with a running mean. This processing was introduced to find non-ambiguous losses of signal which is particularly important for real observations. Signals in the lower troposphere are quite noisy and a unique detection of these events is otherwise ambiguous since the algorithm will also find events which are mainly noise dominated at lower altitudes. The actual reduction in FSI amplitude was then calculated at the altitude of the strong gradient. Variations in the smoothing interval shows no great sensitivity.

Figure 5 shows the resulting reduction in the normalized FSI amplitude for different refractivity gradients at the altitude of the perturbation for a closed loop, fly-wheeling enabled tracking and for an open loop tracking type [Sokolovskiy, 2001b]. We note that the fly-wheeling model was implemented to achieve consistency with the CHAMP observations; the implementation should not be regarded as an accurate model of the "BlackJack" receiver aboard CHAMP, although simulations show very good agreement with actual CHAMP measurements. The required open loop Doppler model is derived from a set of tropical radiosondes. Since closed loop tracking is very sensitive to noise, 50 different calculations were performed for refractivity gradient. Open loop shows only a very small noise impact and we therefore show only one calculation.

Gradients around $-100 \text{ km}^{-1}$ show mean reductions of the FSI amplitude by about 0.5; early loss of signal tracking as well as down-sampling from 1 kHz loop update frequency to 50 Hz output frequency contribute to the observed reduction for a closed loop receiver. Gradients below the ducting criteria of $-160 \text{ km}^{-1}$ generally lead to a strong reduction in signal strength and loss of tracking. The occurrence of early tracking loss is reduced at stronger signal levels; calculations with a signal carrier-to-noise density ratio of 95 dB Hz only show FSI amplitude reductions of up to 30 % for very low gradients. Further simulations showed that this reduction is caused by the data down-sampling process.

A signal carrier-to-noise density ratio of about 55 dB Hz allows successful tracking down to about $-150 \text{ km}^{-1}$ for a closed loop algorithm in this implementation. Placing the refractivity gradient layer at an altitude of 3 km shows that signal reductions exceeding 50 % are already observed at about $-100 \text{ km}^{-1}$ for a signal carrier-to-noise density ratio of 55 dB Hz and a closed loop receiver. A 60 dB Hz value allows tracking up to gradients of about $-150 \text{ km}^{-1}$ at these lower altitudes. Thus successful tracking into the PBL using a closed loop receiver requires higher signal carrier-to-noise ratio densities, at least for very sharp transitions at the top of the PBL. Actual ducting gradients in ECMWF re-analysis data can go down to $-500 \text{ km}^{-1}$, although globally they are close to the $-160 \text{ km}^{-1}$ threshold [von Engeln and Teixeira, 2004].

For the near future it is planned to upload an implementation of the open loop algorithm onto the CHAMP receiver. The impact of refractivity gradients on the FSI amplitude for a simple open loop algorithm is also shown in Figure 5 (right). Open loop tracking is not as affected by noise as closed loop, and tracking can in theory continue even up to very high gradients in refractivity. FSI amplitude reduction found here are caused by the down-sampling from 1 kHz to 50 Hz. In open-loop tracking mode this conversion causes a drop in SNR observations, in particular within multipath regions, since the corresponding 50 Hz inphase and
quadphase correlation sums are obtained by averaging over strongly fluctuating 1 kHz data.

Although Figure 5 gives information on the refractivity gradient where tracking might be lost, no information on the accuracy of the retrieved refractivity profile from the measurement is given. Thus tracking might still be possible, but retrieved refractivity values could be biased with respect to the true profile. In fact, the applied CHAMP processing setup has been derived by studying the introduced refractivity bias for low FSI amplitudes. The retrieved refractivity bias found near the perturbation altitude for closed and open loop tracking is shown in Figure 6 for different noise levels. Closed loop calculations are again averaged over 50 realizations while open loop uses just one. Additionally, a calculation with no receiver present in the simulator shows the ideal case.

Closed loop tracking introduces a bias in the retrieved refractivity, starting already at gradients of about $-60\,\text{km}^{-1}$ for a 50 dB Hz signal carrier-to-noise density ratio. This improves with higher ratios. The standard deviation of the retrieved refractivity profile for closed loop tracking (50 dB Hz) is around 0.3 % for gradients above $-70\,\text{km}^{-1}$ and then shows a strong increase to about 3 to 6 % below; for the 60 dB Hz calculation low standard deviations are found up to about $-160\,\text{km}^{-1}$ and then increases to 6 % (not shown). Open loop allows unbiased tracking to about $-140\,\text{km}^{-1}$ with a 50 dB Hz signal carrier-to-noise density ratio. Higher ratios lead only to small improvements in the retrieved refractivity bias. No pronounced difference is found between the open or closed loop calculation with 60 dB Hz signal carrier-to-noise density ratio for gradients above about $-120\,\text{km}^{-1}$.

The biases found in Figure 6 in open and closed loop tracking for strong super-refraction are explained by looking at the ideal case with no receiver present. Below refractivity gradients of $-160\,\text{km}^{-1}$ critical refraction causes deviation between true and retrieved refractivity [Sokolovskiy, 2003]. Closed loop tracking may introduce an additional bias, it is caused by the imperfect receiver tracking. It starts already at gradients of about $-60\,\text{km}^{-1}$ for a 50 dB Hz signal carrier-to-noise density ratio, this improves with higher ratios. The open loop simulations are close to the ideal case up to about $-150\,\text{km}^{-1}$.

As mentioned above, the Doppler model used in the open loop tracking was derived from a set of tropical radiosonde profiles. The Doppler model could potentially lead to biases in the derived refractivity profile if it deviates too strongly from the actual observed Doppler. We redid the presented open loop calculations with a Doppler model that has no deviation and found very similar results.

4. Data Analysis

Based on Figure 5 and the AWI sonde results shown in Figure 4 one can make an estimate of the number of CHAMP tracking losses that could be caused by strong refractivity gradients. The simulations presented above show that a cer-
Figure 6. Retrieved refractivity bias with respect to the input profiles given in Eq. 3 over different refractivity gradients for closed, open loop tracking, and no receiver. Signal carrier-to-noise density ratio is given in legend.

tain percentage of processed data will show an FSI amplitude reduction larger than 0.5 in the presence of a certain refractivity gradients at a certain altitude where results presented are for an altitude of 6 km. One can get a relationship between the refractivity gradient, the associated altitude, and the percentage of simulations showing a strong FSI amplitude drop at this altitude by repeating these simulations at other altitudes (the impact of these refractivity gradients on the FSI amplitude at other altitudes is similar to the results at 6 km). Calculating the gradients corresponding to Figure 4 found in the AWI data and using this relationship between altitude, gradient, and FSI amplitude reduction, one finds that about 12% (4%) of AWI radiosondes could cause FSI reductions of 0.5 at low (mid) latitudes at altitudes above 3 km.

The mean percentage of CHAMP profiles lost above 3 km within a certain box (as shown in Figure 1, right) at the AWI locations is about 46% (17%) for low (mid) latitude occultations; these numbers are weighted with the number of AWI radiosondes that fell inside a box. Hence these numbers show that only a small fraction of tracking losses can be explain by strong refractivity gradients.

The actual number of CHAMP occultations affected by strong gradients can be estimated by analyzing the FSI amplitude for a sudden drop in strength. We use this characteristics to search in all available CHAMP radio occultation data up to June 2005. In total, more than 200,000 profiles have been processed where CHAMP currently operates in closed loop mode. The operational processing at GFZ stops when the normalized FSI amplitude drops below a threshold of 0.5 to avoid the introduction of too strong refractivity biases, note that here we do not apply this threshold and analyze the full amplitude.

In addition to the smoothing performed in Section 3, only observations were selected where the FSI amplitude dropped by more than 50% over the 500 m vertical interval. This removes the noise present in the FSI amplitude at lower altitudes. Using a different cutoff does not change the general findings of our study. A higher one finds less events while a smaller one finds more events, but noise levels here are also increased.

In FSI processing, the amplitude is given over impact parameter \( a \) or impact parameter over radius of curvature \( a_C \), where \( a_C = a - r_C \) with \( r_C \) the radius of curvature. Often \( a_C \) is also called ray height. Actual altitudes with respect to the earth surface have been approximated with ECMWF re-analysis data and a spherically symmetric atmosphere, using Bouguer’s rule [Born and Wolf, 1980]. The standard deviation of this approximation is generally below 300 m, where sea based occultations show lower values. Although the CHAMP satellite provides refractivity measurements, ECMWF data was used over here to avoid the introduction
of altitude errors caused by erroneous refractivity data calculated from CHAMP observations in the presence of strong refractivity gradients.

Figure 7 shows the positions and the color coded altitude of the events found. In total there are more than 15,000 events, with about 8,600 over land.

The map clearly shows that the algorithm works and finds occurrences of signal loss; most of the events found near the surface are over land, especially over polar regions, e.g. Greenland or Antarctica. Also, mountainous regions show up clearly on the map, e.g. Andes, Rocky Mountains, Himalayas. These events are generally caused by the sudden disappearance of the GPS satellite signal behind the Earth. But additionally there are also a significant number of events found at higher altitude, mostly over the sea in convectively active areas at altitudes up to about 10 km.

The event distribution also has a seasonal oscillation (not shown), especially for land based ones, partly connected to the general flow of humidity with the seasons. The number of events in central Asia and Northern America show a minimum in Northern Hemisphere summer, when higher atmospheric humidity will lead to a degradation of the FSI amplitude at higher altitudes. A similar pattern is found for events near the Andes, where a minimum is found in Northern Hemisphere winter and spring. Also, events over Antarctica peak during Northern Hemisphere winter, indicating that the strong temperature inversions found there primarily in summer affects the FSI amplitude [von Engeln and Teixeira, 2004].

The total number of sea based events found at altitudes above 3 km represent about 13% (21%) of all low (mid) latitude occultations that are lost in these areas (see Figure 1 right). Restricting the events to boxes where also AWI sonde data is available, about 6% (3%) of all occultations within these boxes are affected at low (mid) latitudes (corresponding to about 13% (25%) of all occultations lost above 3 km at the AWI locations). It also agrees fairly well with the estimation made above; our analysis of the CHAMP FSI amplitude will underestimate the number of occultations affected, e.g. noise on the FSI amplitude can make an unambiguous identification difficult.

Although the found events are only a small fraction of the occultation lost at higher altitudes, they can still be used to gain better understanding of the receiver tracking behavior, and also allow to study the impact of atmospheric layers with enhanced water vapor. The altitude distribution of the events is shown in Figure 8 for three latitude bands, separated by land and sea events.

Low latitude events over the sea show a maximum of occurrence around 7 km and less pronounced maxima between 2 km and 3 km. The lower maximum is associated with the top of the PBL, especially with events when the PBL top is at higher altitudes. This can also be seen in Figure 4 where sharp gradients in RH can extend up to altitudes well above the PBL. Hence the PBL top found here corresponds more to the upper range of top PBL values, as also reported in Beyerle et al. [2005] for the AWI radiosonde dataset. The minimum inbetween these two maxima is also visible in Figure 4.

The applied data processing will mainly pick up events that have not entered the lower troposphere where a degradation of the FSI amplitude in the presence of water vapor variations is observed. No events are therefore found near the surface for sea based observations. Land based low latitude events show no distinct altitude dependence, they are found up to 10 km. The amplitude reduction with decreasing altitude due to the water vapor presence masks the PBL which consequently does not show up here.

Mid latitude sea events show similar characteristics although the top maxima is lowered by almost 2 km. This upper maximum is not seen in the radiosonde analysis (Figure 4) which might be caused by the actual locations of the radiosonde observations as opposed to the CHAMP global dataset. Most AWI radiosondes are close to the coast at mid latitudes. There is also an increasing number of events that terminate near the surface, these sea events are often found near the 60° S latitude, an area where very low probability or no ducting was found in ECMWF re-analysis fields [von Engeln and Teixeira, 2004]. Thus atmospheric conditions in this area allow tracking down to the Earth surface. Most land based events near the surface are found in the mountainous regions indicating that the signal was probably terminated by hitting the Earth, although altitudes in Figure 8 range up to 2 km. Two contributions can be identified to these erroneous altitudes: 1. the mapping of FSI amplitudes to altitudes using ECMWF data; 2. the actual position of the tangent point, where a few kilometers misplacement are especially important in mountainous regions. The latter contribution is the major error source. In fact, obtained tangent points from GFZ and another CHAMP processing center show deviation [von Engeln, 2006] where the actual cause is currently investigated.

High latitude sea events do not show a very distinct maximum at higher altitudes. They are mainly found near the surface. Land based events here are similar to the mid latitude ones although the high altitude terrains in polar regions lead to an earlier increase of found events around 3 km.

The observations (by Mapes and Zuidema [1996]; Brown and Zhang [1997]; Zuidema [1998]; Johnson et al. [1999]; Newell et al. [1999]; Thouret et al. [2000]) introduced in Section 2.2 agree with our findings. The convectively ac-
**Figure 7.** CHAMP altitude above earth surface where sudden signal loss was found.
Figure 8. Number of occurrences at particular altitude separated by latitude (30° bins) and land sea events. Bin size 0.35 km.
tive zones in the ITCZ and SPCZ show a reduced number of CHAMP observations in the lower troposphere (see Figure 1, right). Also, signal losses are frequently found in the tropics around 7 km, with a clear distinction between land and sea based events (see Figure 8) and a layer with lower occurrences between about 3 km to 4 km. The seasonal behavior in our data also confirms a minimum number of events for northern latitude winter at altitudes above about 4 km. Although there is no clear spring maximum for tropical Asia since data is too noisy.

5. Conclusion

Radio occultation measurements using GPS signals offer high vertical resolution profiles of atmospheric refractivity from about 35 km down to the surface, which can be converted into profiles of temperature and pressure. When the method was first proposed, it was thought that observations into the PBL would generally be possible. Recent instruments, e.g. CHAMP, however show that quite a significant number of observations are lost well above the PBL, especially at tropical latitudes, e.g. in the Intertropical Convergence Zone (ITCZ).

One possible cause for signal tracking loss at these higher altitudes are strong refractivity gradients in the atmosphere. It is shown that these strong refractivity gradients can theoretically be generated by changes in the water vapor concentration over sharply defined layers. Generation is possible up to altitude of about 6 km to 7 km in the tropics for gradients of \(-100 \text{ km}^{-1}\), the generation of ducting (or super-refraction at \(-160 \text{ km}^{-1}\)) conditions is only possible up to about 5 km to 6 km, under the assumption that the air is not super-saturated. Radiosonde measurements show that decreases in the water vapor concentration over a thin layer can be up to 40% in relative humidity at altitudes of 6 km.

Wave optics simulations using fly-wheeling CHAMP like receiver settings were used to determine the influence of such refractivity gradients on the signal tracking. With settings typical for low latitudes, it was found that sudden drops in the FSI amplitude are found in the presence of strong refractivity gradients and that they can already be generated by gradients around \(-100 \text{ km}^{-1}\). The simulations showed in addition that such reductions in FSI amplitude are also caused partly by down-sampling to the receiver output frequency of 50 Hz. Furthermore, loss of signal tracking will also happen when the GPS satellite sets behind the earth.

We then searched in CHAMP data for these sudden drop in signal strength, as found in the FSI amplitude simulation. With our search algorithm, about 7% of all CHAMP observations showed a sudden drop in FSI amplitude. At polar latitudes, this was mainly caused by the setting of the satellite. Also, mountainous regions stand out. For low and mid latitudes, events found over land and sea show different signatures. Land events show generally only very little altitude dependence, although near the surface mid latitude events increase. But both altitude bands show an increased number of events near 7 km (low latitudes) and 5 km (mid latitudes) and peaks near the PBL. These events at higher altitudes were attributed to atmospheric layers with sharply defined tops, causing larger refractivity gradients and thus tracking loss.

Such atmospheric layers around the 0°C level (around 5.5 km) are more stable and control the depth of cumulus congestus clouds. The increased humidity, along with radiative cooling, at the top of the cloud leads to strong refractivity gradients and thus signal tracking loss. Results shown here also show the prominent minimum in cloud cover and relative humidity, corresponding to a minimum in the number of events found between about 3 km and 4 km, and an increased number of events at the top of the PBL, at about 2.5 km.

We draw two main conclusions from this study of CHAMP radio occultation data:

1. The high vertical resolution of radio occultation data allows to study thin atmospheric layers in the middle troposphere. The altitude where cumulus congestus cloud tops are present was successfully identified in CHAMP FSI amplitude data. Thus radio occultation instruments offer a unique opportunity to study this important area of the troposphere, where changes in the humidity profile will affect the radiative and latent heating profile and hence the dynamics. Especially future instruments with better tracking capabilities should enhance our understanding of processes in the middle to upper troposphere.

2. The successful tracking of these events without signal loss requires signal carrier-to-noise density ratio of about 55 dB Hz for a closed loop algorithm in our implementation, where the CHAMP instrument value is about 50 dB Hz for low latitudes. Even higher values are required to track successfully into the PBL, where 60 dB Hz is generally sufficient. Open loop algorithm are also not as strongly affected by these thin atmospheric layers in terms of introduced refractivity biases. Comparisons of open and closed loop calculations show similar introduced refractivity biases for 60 dB Hz noise values up to refractivity gradients of about \(-120 \text{ km}^{-1}\). The use of open loop tracking should also allow to investigate thin atmospheric layers directly using the refractivity profile.
Future work will look at possible signal loss caused by super saturation with respect to ice, or stratospheric intrusions at mid latitudes. The ever increasing dataset of the CHAMP instrument also offers the opportunity to study seasonal and diurnal effects in these strong refractivity gradients and thus in the vertical humidity distribution.

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