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The Cryosphere Discussions

Interactive comment on "Density assumptions for converting geodetic glacier volume change to mass change" *by* M. Huss

M. Huss

matthias.huss@unifr.ch

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The comments by Reviewer #1 address some important points, mainly related to the firn densification model, and were very helpful to revise the manuscript. In response to this review, I have added detailed discussions and performed additional model experiments. As requested, a table now provides details on the observed firn density profiles in different mountain ranges. The set of profiles included has been extended.
My responses to the review (text written in *italic*) are given, including proposed changes to the text of the paper (in quotation marks).

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1. The HL-model distinguishes two phases of firn compaction: an initial 'rapid' *compaction phase, where steady-state firn compaction is not a function of accumulation, and where compaction is believed to be primarily controlled by grain settling and packing. Below a density of 550 kg m^{-3} this is followed by a second stage of compaction, where grain growth and sintering are the dominant processes, until pore close-off. These different phases are not described in the manuscript, which can be improved for the sake of completeness. Equation 6 describes the second stage of firn compaction in the HL-model, which means that the first stage of compaction is essentially neglected. However, since firn density at the surface is set to 520 kg m^{-3} (a high value, see below), the initial fast compaction phase would only occur in a very thin layer, so the difference with the HL-model is only minor. Nevertheless, a comment on this issue would be welcome.*

Thanks for this comment.
The Herron and Langway (1980)-model has been developed for ice sheet firn compaction for which the densification regime strongly differs from temperate or polythermal mountain glaciers mostly due to large differences in typical accumulation rates and firn temperatures. The initial density of about 500 kg m^{-3} (see further discussions below) refers to the density of one-year old snow (i.e. newly formed firn) that are characteristic for mountain glaciers. Snow densification (i.e. the first year of compaction) is not considered here; the firn densification model is initialized with the density of snow/firn one year after deposition. Therefore, I assume that the first stage of compaction (below a density of 550 kg m^{-3} after Herron and Langway) mostly takes place during the first year within the snow layer, actually before the firn model is initialised. A compressed version of this discussion is included in the paper.

”The HL-model specifies two phases of compaction in dry firn. Here, the first densification phase (according to Herron and Langway (1980) for ρ <550 kg m^{-3}) is not considered as it is assumed to take place in the first year after snow deposition and to only have a minor importance for mountain glaciers.”

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2. Figure 2 illustrates that the original HL-model does not result in a good match between observations and simulated density. To solve this, the compaction rate has been increased, to a tuned value of 0.11 0 m$^{-0.5}$ a$^{-0.5}$. Though effective, this adjustment seems not physically realistic: using k_1 = 575 exp(–21400/(RT)) this means that the accompanying temperature is ≈300 K, i.e. not possible in a firn layer. As the HL-model has been widely recognized to result in a quite good agreement on dry compaction, it is more likely that instead of the dry compaction, it is the refreezing component that should be adjusted to make the model in line with the observations. How the amount of refreezing is calculated is not extensively addressed. It seems more reasonable to increase the influence of refreezing, to get an improved match between simulated and observed density profiles.

This is an important point to be addressed. In fact, I did not simply tune k_1 but assumed T=273 K for temperate firn and calibrated the factor f = 1380 (Herron and Langway (1980) recommend f = 575) in the equation
$k_1 = f * exp(-21400/(RT))$.
As k_1 is constant for temperate conditions (as stated in the TCD paper) I have lumped all these factors together for the sake of simplicity and only stated the value of k_1. My calibration is thus not unphysical for temperate glaciers, but provides a different estimate for the factor f. f is the 'tuning parameter' in Herron and Langway (1980). They derive f based on a linear regression of observed densification in a dry-compaction regime on the ice sheet (see their Fig. 1). It seems plausible that this factor might change for temperate conditions that experience different processes of firn densification. In several studies, the compaction of temperate firn has also been described using viscous deformation (e.g. Ambach and Eisner, 1983, AoG, Vimeux et al., 2009) instead of an empirical or physical firn compaction model.

In my model, refreezing ($R(t)$ in Eq. 5) is determined by estimating an end-of-winter temperature profile providing a negative heat reservoir which is completely used to refreeze melt water (and thus to densify the firn layers). This approach is now better described in the paper. Seasonal temperature variations (i.e. the penetration of cold winter temperatures into the uppermost firn layers) can be simulated using heat conduction or be directly measured (e.g. Hooke et al, 1983; Greuell and Oerlemans, 1989; Haeberli and Funk, 1991).

If f is set to 575 (according to Herron and Langway, 1980) as suggested in the reviewer's comment and the firn densification is tuned with refreezing only, unrealistically low end-of-winter firn temperatures (i.e. far beyond the range of direct measurements, see Hooke et al, 1983) are required to reproduce the observed rates of firn densification. It is thus concluded that for polythermal and temperate firn, a re-calibration of the parameters of the original HL-model is justified.

The tuned model parameters of the TCD paper were thus not changed, but the revised paper discusses this issue, provides the full definition of the parameter k_1, and gives more details (and references) on the calculation of refreezing.

"The term k_1 is constant for temperate conditions and is defined as

$$k_1 = f \cdot exp\left(\frac{-21400}{RT}\right),$$

where R is the gas constant, T the firn temperature in K, and f a factor that was empirically determined by Herron and Langway (1980). In this study, f is used to tune simulated to observed firn densification (see below).

In order to keep the model simple (not requiring climate data input) and location-independent, firn densification due to refreezing $R(t)$ (Eq. 5) is roughly approximated by assuming an end-of-winter firn temperature profile that linearly increases from $-5\,^{\circ}$C at the surface to 0 °C at a depth of 5 m. This profile corresponds to the typical penetration depth of winter air temperatures (e.g., Hooke et al., 1983; Greuell and Oerlemans, 1989) and defines a negative heat reservoir for refreezing melt water. For each firn layer, r is obtained by prescribing complete latent heat exchange. Total refreezing $R(t)$ after t yr is calculated

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as $R(t) = R(t-1) + r$. ”

”Application of the original HL-model (including an additional term for refreezing, Eq. 5) for a location with mean annual accumulation rates equal to those observed by Ambach and Eisner (1966) with f=575 (Eq. 7, following Herron and Langway (1980)) results in too slow firn compaction (Fig. 2). The parameter f=1380 is optimized to match density profiles obtained from the firn cores (Fig. 2). ”

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3. A total of 12 measured firn density profiles, probably from very different climatic settings, are lumped together to provide a mean density profile including variability. Although the author argues that a good comparison with the calibrated firnmodel indicates that this empirical method can be applied to describe firn compaction for a wide range of mountain glaciers, this merely seems a matter of choosing the right locations. A table with precipitation and mean annual temperature values for these locations would support this claim.

Such a table is included in the revised paper. Based on the additional studies pointed out during the review process, the set of firn density observations on mountain glaciers was extended to 19 cores in total.
Data on air temperature and annual precipitation as requested by the reviewer were however not available for all firn density measurements. The table provides an estimate of the mean accumulation rate, and a qualitative statement about the firn temperature regime (temperate, polythermal). The mean density of the uppermost 10 m was also evaluated and is given in the table (see below).

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4. The value of the density of fresh snow (520 kg/m3) is very high, also with respect to the composite of observations (12 firn density curves). It is also strange that the calibrated model results in a firn density of ≈600 kg m^{-3} at the surface. Why is this

Table 1. Firn density data compiled from studies on temperate and polythermal glaciers in different mountain ranges. For each firn core the average density of the first 10 m $\overline{\rho_{10}}$ (in kg m^{-3}), the approximate mean accumulation rate b (m w.e. a^{-1}), and a qualitative statement about the thermal regime is given. Some studies provide more than one firn core (n_c), and values refer to an average.

Reference	Location	n_c	$\overline{\rho_{10}}$	b	type
Ambach and Eisner (1966)	Europ. Alps	1	700	1.2	temp.
Oerter et al. (1982)	Europ. Alps	3	600	≈1	temp.
Sharp (1951)	Western Canada	1	650	≈1.5	temp.
Zdanowicz et al. (2012)	Arctic Canada	2	560	0.3-0.6	polyth.
Kreutz et al. (2001)	Tien Shan	1	650	1.3	polyth.
Suslow and Krenke (1980)	Pamir	1	620	≈1	polyth.
He et al. (2002)	Himalaya	1	640	0.9	polyth.
Matsuoka and Naruse (1999)	Patagonia	1	620	2.2	temp.
Shiraiwa et al. (2002)	Patagonia	4	550	5-15?	temp.
Pälli et al. (2002)	Svalbard	1	510	0.4	polyth.
Nuth et al. (2010)	Svalbard	3	510	≈0.5	polyth.

not 520 kg m^{-3}?

The value of 520 kg m^{-3} does not refer to the density of fresh snow but to the 'density of new firn' (P225, L8). Per definition, firn is at least one year old snow, so this is the snow density typically measured on mountain glaciers one year after deposition. Based on the average density of the first annual firn layer of the 19 firn cores compiled in the additional table, a value of $\rho_{\text{firn},0}$= 490 kg m^{-3} is defined, which represents a minor change compared to the TCD paper. The basis of this value is described in the revised manuscript.

"... , and $\rho_{\text{firn},0}$ is set to 490 kg m^{-3} according to the average density of the uppermost annual firn layer compiled from 19 density profiles on mountain glaciers in different climatic regimes (Table 1). ”

Thanks also for the second remark. The figure has been re-drawn, and the starting density $\rho_{\mathsf{firn},0}$=490 kg m^{-3} is indicated. Equation (5) makes it clear that the density at time t=1 yr does not correspond to $\rho_{\mathsf{firn},0}$.

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Firn compaction rate is largely determined by temperature. In this study, only experiments with a change in surface mass balance forcing are carried out. It would be a valuable addition if firn compaction effects induced by temperature changes would also be addressed.

The firn densification rate is strongly determined by temperature for a dry-compaction regime on the ice sheets and in polar regions. I assume that the temperature-dependence of the densification rate will be much smaller for temperate firn of mountain glaciers that are the focus of this study.

The firn densification model used here is crude and does not explicitly include a temperature forcing (no air temperature data are needed to drive the model). The refreezing term $R(t)$ in Eq. 5 however depends on temperature and was used for an additional sensitivity test. This experiment prescribes a doubling of the refreezing rate and thus allows judging (rather qualitatively) the impact of a changing temperature forcing on $f_{\Delta V}$ as it might occur in the case of a transition from cold to temperate firn (e.g. Zdanowicz et al., 2012). The experiment is included in the Discussion section and results are given.

” The sensitivity of $f_{\Delta V}$ on the climate regime, i.e. the surface mass balance distribution, and potential changes in characteristic compaction rates with higher temperatures, was analyzed (i) by reducing balance gradients by a factor 2 for all experiments, corresponding to more continental conditions, and (ii) by increasing the amount of refreezing $R(t)$ (Eq. 5) by 100%. The approach to model $f_{\Delta V}$ is relatively insensitive even to these major changes in firn compaction. For reduced mass balance gradients, the average conversion factor is 10 kg m^{-3} below the reference results (excluding Experiment III). Doubling the refreezing rate $R(t)$ causes an increase in average $f_{\Delta V}$ by $+20$ kg m^{-3}. More refreezing is expected in a warming climate for cold or polythermal glaciers (e.g. Zdanowicz et al., 2012) . For temperate mountain glaciers higher temperatures will however rather reduce $R(t)$. "

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Page 222, line 17: Helsen et al. (2008) showed that Antarctic accumulation variability is the main driver behind observed elevation changes, not necessarily density changes.

Right. Firn density variations replaced with firn depth variations.

"Helsen et al. (2008) show that surface elevation changes in the interior of Antarctica are mainly due to firn depth variations and can not be interpreted as a mass change."

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Page 228, line 22: 'It is remarkable that already with a minor change ...' Why is it so remarkable that even short-term variability causes volume changes that occur with a near-surface density of the firn? In fact, this is one of the major points of this paper: especially when perturbations are short, these are accompanied with values of f$_{\Delta V}$ that strongly deviate from the ice density. Consider revising this sentence.

Reformulated.

"Already with a minor change in mass balance, as in the first year after the spin-up phase (ΔELA $= 5$ m), $f_{\Delta V}$ significantly diverges from 900 kg m^{-3}."

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Fig. 4: it is unclear what the different grey lines exactly represent. In the text this is only briefly described, it has to do with different elevation ranges, but the details are not clearly presented.

Additional details provided in the figure caption.

Caption Fig. 4:
"Thin grey lines refer to model runs using different glacier size (elevation ranges 300–2000 m), solid lines show the mean of all experiment simulations for a positive/negative shift in the ELA. ..."

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Page 230, line 25: values between –550 and 6500 kg m^{-3}? I cannot find this value in the figure.

Good point. Figures 5a/b only show a part of the series (1990-2005). These numbers were derived from the complete series of annual $f_{\Delta V}$ for both glaciers (n=92). This is clarified.

"Evaluation of the complete series of annual conversion factors for both glaciers (n=92) shows that $f_{\Delta V}$ range between minimum/maximum values of -500 and 6500 kg m^{-3} for mass balances B_a of -0.2 to $+0.2$ m w.e. a^{-1} ..."

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A number of sensitivity experiments are mentioned. It is not clear which of these are also illustrated in the results, and how they can be recognized (Figure 4, 5). For example, it is mentioned that alternative values for the slopes along the elevation

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range are used, but I could not find this back in the figures. This can be improved.

No results of sensitivity experiments are shown in Figures 4 and 5. Displaying all of these results in the figures would make them difficult to read and dilute the main points. Results of all sensitivity test are only given in the text. This is clarified in the revised paper.

"Several sensitivity tests were performed to investigate the robustness of calculated $f_{\Delta V}$ on changes in the synthetic glacier geometries and simplifications in the modelling of firn densification. The model was run with idealized mass balance forcing for all experiments (Fig. 3). Differences in the computed conversion factor were compared to the "reference" results (shown in Fig. 4 and Table 2) and results are discussed hereafter. "

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The sensitivity to choices in the parameters in the firn compaction model was tested. However, the choice for a value of 520 kg m^{-3} for the surface density is not tested, while this is a very high value. Accumulating fresh snow can have values below 300 kg m^{-3}. This generally will result in much larger deviations of f$_{\Delta V}$ from the ice density.

Good suggestion. An additional sensitivity experiment addresses the uncertainty due to the assumed initial density. This value strongly differs between regions. Whereas in cold environments, densities of new firn (i.e. one-year old snow) might be quite low (300-400 kg m^{-3} on Svalbard glaciers, e.g. Pälli et al., 2002), they can easily reach $>$600 kg m^{-3} at temperate high-accumulation sites (e.g. Ambach and Eisner, 1966). The impacts on calculated $f_{\Delta V}$ are relevant. Minimum and maximum initial firn densities are taken from the 19 profiles with firn density data.

" Varying the initial firn density $\rho_{\text{firn},0}$ between the minimum and the maximum of the observed values (Table 1) leads changes in $f_{\Delta V}$ of -31 and $+26\,\text{kg}\,\text{m}^{-3}$, respectively. "

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Page 233, line 13: Here it is suggested that a linear fit is calibrated with the observations, but this cannot be seen in Figure 2.

The reference to Fig. 2 is now omitted.

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In the conclusions it might be worthwhile to repeat the explanation of how values of $f_{\Delta V}$ can be higher than the ice density.

Done.

"For the particular case of strong changes in mass balance gradients together with limited mass gain/loss, $f_{\Delta V}$ can however also be systematically higher than the ice density as opposite signs of elevation changes in the accumulation and the ablation area can compensate for each other in terms of volume. "

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