Groundwater springs in the Hedmarksvidda mountains related to the deglaciation history

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Numerous groundwater springs in the Hedmarksvidda area are related to ice marginal deposits from the Early Holocene. The spring aquifers are small and consist of coarse glacial sediments. The bedrock consists of Late Precambrian conglomerates, sandstones and shales. Oxygen isotope composition indicates a short residence time; a few weeks to a couple of months. The groundwater has low ion contents, due to the short transit time and because the glacial sediments of the spring aquifers are formed mainly by material from conglomerates and coarse-grained arkosic sandstones. The vegetation surrounding the springs is typical for cold water with a low nutrient content. Mountain farms and cottages are frequently located near springs, and sheep and other herbivorous mammals are attracted to the areas with spring herbs. The discharge from the springs controls the surface stream base-flows, which are significant and stable during the winter time, but much less than what is measured during the period May—September. The mountain springs are important for the local hydrology and for maintaining the ecosystem of the area.

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Groundwater springs in mountain areas are important for the local ecosystem. The groundwater temperature has a small seasonal variation compared to surface water temperature. In winter, the water is warm compared to the surrounding snow fields, and they may be the only open local water sources. In the summer, the groundwater springs are cooler than other sources of surface water.

Mountain farms have been established in the vicinity of groundwater springs, where it has been possible to cool the milk during the summers. During the entire year the open spring outlets are a source of drinking water for animals and people.

In most mountain areas groundwater springs are found where the bedrock is fractured and the overburden is thin. However, in mountainous areas springs are quite frequent in Quaternary sediments too (Nordhagen 1943). The upper part of Åstdalen in Hedmark, southeastern Norway (Fig. 1), in the Hedmarksvidda mountains, is an area where groundwater springs in thick Quaternary glacial sediments are particularly frequent.

The aim of this paper is to describe the typical characteristics of such mountain spring aquifers and their relationship to the glacial history.

Site geology

The bedrock in Åstdalen is formed by the Late Precambrian Brøttum Formation. In the north, black shales and greywacke sandstones dominate, whereas in the south, arkosic sandstones and fine-grained conglomerates are most common (Fig. 1). The bedrock is moderately fractured with more intensively frost-weathered surfaces occurring above an elevation of 1100 m. The overburden is dominated by till and reworked till material. The glaciogenic sediment cover has an average thickness of several metres, which is thick within a Norwegian context, and bedrock exposures are rare.

Position of the main spring horizons in Åstdalen

The upper spring horizon

The upper and most marked spring horizon is found along prominent terminal moraines and hummocks at an elevation from about 950 m in the south up to 1100 m in the north. This level marks the ice margin during a phase of the last deglaciation (Fig. 2). Radiocarbon dating of gyttja from Lille Møklebysjøen just north of the main moraine ridge in Godlidalen indicate a Preboreal age (T-9927A, 9650 ± 140 BP; T-9927B, 9580 ± 130 BP) (Groseth 1992). At that time the glacier almost completely filled the main valley and only the highest summits along the tributary valleys were ice-free.

During this phase, the active glacier deposited a compact basal till. Flow till and glaciofluvial sediments were deposited laterally (Fig. 3A). Along ice-free valley sides, previously deposited till was exposed to solifluction and partially sorted by water. An intensive frost weathering took place where the overburden was thin. When the ice melted, some of the deposits along the valley sides flowed downslope and partly covered the upper part of the compact basal till. Along steep parts of the main valley and its tributaries, terminal moraine ridges were nearly completely destroyed by secondary flow processes. Only in the flat-lying areas were the terminal moraine ridges
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preserved. The laterally deposited sediments and resedimented tills along the upper part of the valley sides are characterized by a lower degree of compaction and a lower content of fine-grained material than the compact basal till. However, the whole cover of Quaternary sediments is dominated by diamictons, and the upper, hummocky sediment complex would previously have been classified by many glacial geologists as ‘ablation till’, in contrast to the lower lying compact basal till.

Single groundwater springs and groups of springs are found at the boundary between the more permeable sediments and the underlying or lower lying compact basal till (Fig. 3B). They can be classified as contact springs. The most marked horizons are observed in Godlidalen and along the mountains of Grávola, Lyngkampen and Høgfjellet (Fig. 2). In these areas numerous springs are found side by side, with more diffuse seepage zones between them.

The water from the spring horizons gives enough moisture to form downstream peatland areas, which cover most of the basal till.

The lower spring horizon

Many single springs are found at a level of 900 m in the northern part of the main Åstidalen valley. As for the upper spring horizon, they are concentrated at a level where there are marked lateral glacigenic deposits. At the time these were deposited the ice nearly filled the main valley, while the tributary valleys and mountainous areas were ice-free. Terraces of glaciofluvial material or dia-
micron form the main spring acquifers. They occur mainly as single springs and not in groups as in the upper horizon. They are of great importance for the siting of mountain farms (Fig. 2), which are located to a large extent on the lateral terraces. The streams that are fed by groundwater from the springs at this level are utilized frequently by farmers and tourists for water supply.

Groundwater budget in Godlidalen

Godlidalen is a tributary valley to the Åstdalen in the eastern part of the Åsta catchment. The eastern valley side of Godlidalen is particularly rich in springs (Haldorsen 1991; Haldorsen et al. 1992) belonging to the upper spring level (Figs. 2 and 4). More than 80 springs with more diffuse seepage zones between them are found within a distance of less than 3 km. The springs are believed to be representative of the upper spring level as a whole with respect to origin and aquifer type.

Discharge was measured each month during 1989–1991, and continuously during shorter periods in the summers of 1988 and 1989 at one of the most prominent springs in Godlidalen (Fig. 4, spring A). The maximum value measured is 71 l s\(^{-1}\). The discharge of each separate spring in Godlidalen is thus rather small. However, the high frequency of springs produces a high total discharge.

To provide an estimate of the groundwater budget of the spring aquifers in Godlidalen, we consider the groundwater recharge \(R\) to be equal to the discharge \(D\) from the springs. The transport down to deeper bedrock aquifers is thus assumed to be insignificant.
compared with the direct discharge of water through the spring outlets.

The area above the spring horizon up to the water divide \((A)\) is 3.9 km\(^2\) (Fig. 5). The annual precipitation \((P)\) is on average 1100 mm year\(^{-1}\). If we consider the evapotranspiration \((E)\) to be 20\%, the groundwater recharge plus the surface drainage \((R + S)\) per unit area is \((P - E)\), which is about 900 mm year\(^{-1}\). The total annual amount of water in the area above the spring horizon is:

\[
R + S = A \times (P - E) = 3.5 \times 10^6 \text{ m}^3 \text{ year}^{-1} = 0.1 \text{ m}^3 \text{ s}^{-1}
\]  

The maximum value of groundwater recharge \((R_{\text{max}})\) is defined by \(S = 0\), i.e. by considering the surface drainage as insignificant. The true recharge \(R\) must be lower than this:

\[
R < R_{\text{max}}(S = 0) = A \times (P - E) = 0.1 \text{ m}^3 \text{ s}^{-1}
\]  

Haldorsen et al. (1992) calculated the discharge \((D_{\text{min}})\) of groundwater from the springs in Godlidalen during the lowest winter flow conditions measured during the period from 1989 to 1991. The result was:

\[
D_{\text{min}} = 6.3 \times 10^5 \text{ m}^3 \text{ year}^{-1} = 0.02 \text{ m}^3 \text{ s}^{-1}
\]

Normally, the winter discharge is higher than this, and during summer the discharge is generally considerably
higher than during winter. The true discharge \((D)\) must be much higher than \(D_{\text{min}}\).

\[
D = D_{\text{min}} = 0.02 \text{ m}^3 \text{ s}^{-1}
\]  
(4)

The true discharge, \(R\), must be somewhere between \(R_{\text{max}}\) and \(D_{\text{min}}\). This gives:

\[
D_{\text{min}} = 0.02 \text{ m}^3 \text{ s}^{-1} < R < 0.1 \text{ m}^3 \text{ s}^{-1} = R_{\text{max}}
\]  
(5)

The true value of \(R\) is probably not much lower than \(R_{\text{max}}\), because the surface drainage \(S\) above the spring horizon is quite small except during the snow melt period.

Returning to the groundwater spring A (Fig. 4), the average discharge has been calculated to about 21 s\(^{-1}\). this gives an annual discharge of 63,000 m\(^3\) year\(^{-1}\). Divided by the annual recharge of 900 mm year\(^{-1}\), the area from which groundwater discharges in spring A should be about 70,000 m\(^2\) or 0.07 km\(^2\). This means that each spring has a relatively limited recharge area.

**Transit times**

The content of the heavy oxygen isotope \(^{18}\text{O}\) compared with the lighter \(^{16}\text{O}\) in precipitation is dependent on the history of the water vapour, which lasts from the time it evaporates from the ocean until it condenses and forms precipitation. The content of \(^{16}\text{O}\) in precipitation decreases with distance from the coast, with decreasing temperature and with increasing elevation. In the interior of Scandinavia the summer precipitation is significantly richer in \(^{18}\text{O}\) than the winter precipitation (see Lindström & Rodhe 1986; Rodhe 1987). After percolation down to the groundwater zone has started, the ratio of \(^{18}\text{O}\) and \(^{16}\text{O}\) is virtually constant. The characteristics of the original precipitation are maintained. Consequently, the recharged winter precipitation has a lower \(^{18}\text{O}\) content than recharged summer precipitation. Prior to the study in Åstdalen, oxygen isotopes had not been applied in hydrogeological studies in Norway, and except for the studies by Lindström & Rodhe (1986) and Rodhe (1987), very little was known about oxygen isotope variation in different water types in different catchment areas. However, knowledge of seasonal variations in isotopic composition of precipitation seemed very promising for using oxygen isotopes as groundwater tracers in small aquifers.

There are also short-term variations in the oxygen isotope content of precipitation (Lindström & Rodhe 1986), but these variations are not expected to apply to groundwater aquifers because of dispersion in the unsaturated and the saturated zone. In groundwater springs, which are confluence areas for different groundwaters, a significant mixing occurs, and only seasonal variations should be expected.

In Godlilden, the content of \(^{18}\text{O}\), expressed as \(\delta^{18}\text{O}\) (Craig 1961) has been measured in samples from springs A, B and C (for locations see Fig. 4), as well as the precipitation during the years 1989–1991 (Fig. 6). Also, the amount of precipitation and the discharge of Skvaldra and spring A have been measured (Figs. 4 and 7).

Some of the main characteristics of the three groundwater springs are described and discussed below (Haldorsen et al. 1992).

Spring A has marked and rapid variations in discharge and a low ion content, spring B shows smaller variations in both discharge and chemical composition, and spring C has small discharge variations and a relatively high and stable ion concentration. From chemical composition and observations of discharge it is concluded that the residence time of the groundwater is shortest in spring A and longest in spring C.

During the years 1989–1991, the winters were abnormally mild, with low amounts of snow (Fig. 7A). However, even during these three years seasonal variations in the \(\delta^{18}\text{O}\) were observed in the precipitation as well as in the groundwater springs (Fig. 6A–D). Seasonal maxima and minima are seen in spring A and spring B, and are most distinct in spring A. In spring C, the seasonal variation is small with no marked peaks in the oxygen isotope curve. Calculation of the groundwater residence time will be illustrated from the snow melt period in 1989 and the summer/autumn of 1990, from which we have the most complete data.

In 1989 there was an early snow melt. The main spring flood in the Skvaldra stream occurred over 10 days, from 10 April to 20 April (Fig. 7B). Some snow still remained in the field in the beginning of May but it had disappeared before 10 May. The discharge was low at the beginning of May, indicating that most of the snow had melted. In groundwater spring A the \(\delta^{18}\text{O}\) values were relatively high in April compared with May, and the lowest \(\delta^{18}\text{O}\) values were observed at the beginning of May (Fig. 6A). The lowest values indicate that the maximum meltwater had reached the spring A outlet. When this is compared to the discharge of Skvaldra it can be concluded that the average transit time of the meltwater was not more than one month before its appearance at spring outlet A. The short transit time is also shown by the discharge curve of spring A (Fig. 7C), which had its highest values in the middle of June. The sampling frequency is too low to calculate a more exact average transit time.

In 1990 there was a high precipitation rate in June–August, whereas September was relatively dry (Fig. 7A). The \(^{18}\text{O}\) content of the precipitation was high during the same months, but decreased from September to October and dropped significantly in November (Fig. 6D). The \(\delta^{18}\text{O}\) value in spring A increased until the middle of November. If the highest \(^{18}\text{O}\) values in the spring reflect the transit of the summer precipitation, the transit time must be about two months. It cannot be less than about one month because the maximum for \(^{18}\text{O}\) occurred about one month later in the spring than in the precipitation. The indicates a residence time for the summer/early autumn precipitation of one month or longer.
Fig. 6. The variation of oxygen isotopes in springs A, B and C, respectively, (location: see Fig. 4), and D precipitation in Godlidalen in 1989–1990.
As mentioned above, the minima and maxima for spring A were connected respectively to the snow melt and the summer precipitation of the same years. This suggests a residence time of less than one year. If the residence time was longer than one year, a certain amount of discrepancy could be expected between the timing of the maxima and minima of the two springs given their different storage capacities. The fact that the extremes occur at the same time in both springs thus indicates a short residence time, which does not allow for ‘smoothing’ of the sharp seasonal signals. In spring C, increasingly high $^{18}$O values occurred until February 1991 (Fig. 6C), indicating a residence time of several months (Fig. 6C). However, even for spring C, the seasonal variations in the precipitation are reflected in the spring, and the oxygen isotopic composition proved to be a useful parameter in the calculation of transit time.

For all three groundwater springs, the residence time was short, less than one year, and we believe this to be representative for all the mountain springs in Åstdalen. The best evidence for this conclusion is that none of the remaining measured springs have an ion concentration which exceeds that of spring C. If we regard the residence time to be the most important factor influencing the ion concentration, spring C must have a relatively high residence time compared to all the other springs.

The residence time of the summer precipitation in 1990 thus seems to be somewhat longer than the residence time of the meltwater in spring of 1989. A short residence time after the snow melt is probably a general feature, because the unsaturated zone has a high water content at the end of the winter and infiltrated meltwater percolates relatively quickly down to the groundwater zone. In the summer, the situation is more complicated, because the unsaturated zone may have a high or a low water content dependent on the amount of precipitation during the previous period. In addition, much of the water is part of the evapotranspiration process and never reaches the groundwater zone. The transit time for the late summer and autumn precipitation will therefore depend on the field situation earlier in the summer.

The calculation for spring A indicates that the average transit times are not longer than a few weeks. For spring B, the variation in oxygen isotope content is about the same as for spring A; in the spring and early summer of 1989, as well as the summer of 1990, the $\delta^{18}$O maximum occurred at about the same time as at spring A (Fig. 6B).

Groundwater chemistry

The spring water belongs to the bicarbonate facies, in which bicarbonate is the dominant anion (Fig. 8). Ca$^{++}$, Mg$^{++}$, Na$^+$ and K$^+$ are the dominant cations. The springs generally have a low ion content compared with groundwater from wells drilled in bedrock of the Brettum Formation at Lillehammer (Fig. 1). The ion content is also low compared to groundwater with a residence time of more than one year in tills within the Åsta catchment (Fig. 6; Englund 1983, 1986; Haldorsen et al. 1992). This clearly illustrates that the water discharging into the springs has a shorter transit time and less contact with the mineral material than the other groundwater types (see also Englund 1983). There are no differences in chemistry between springs in the northern part of the area, where greywacke sandstones alternate with shales, and the southeastern part of the area, where fine-grained conglomerates and arkosic sandstones dominate (Fig. 1). The spring water originates mainly from coarse Quaternary deposits along the mountain sides, where the underlying bedrock in the north is dominated by sandstones. The shales are usually found in the valley bottom and depressions and do not play an important role in determining the spring chemistry.
Fig. 8. Chemical composition of groundwater springs in Godlidalen in 1991 and data for other springs in Åstdalen in 1977. For comparison, Bedrock well 1 drilled in the greywacke-shale area at Lillehammer; Bed rock well 2 drilled in the arkose-conglomerates area east Lillehammer; and G.w in deep till, groundwater with a long residence time in tills in Godlidalen.

Vegetation

The groundwater spring temperature is 2–4°C. The combination of low temperature and low ion concentration gives a typical spring vegetation, which was classified by Nordhagen (1943) as a \textit{Mniobryum–Epilobium horne}manni (Norwegian: Kaldmose-kildemjølke) association. It is characterized by a low number of species. \textit{Epilobium alsinifolium}, \textit{E. hornemanni} (N: setermjølke) and \textit{Cerastium trinum} (N: brearve) are very characteristic herbs, while \textit{Mniobryum albicans} var. \textit{glaciale} and \textit{Philonotus fontana} (N: kaldemose) form the blanket of mosses that is so typical of the springs (Fig. 9). \textit{Poa pratensis alpigena} (N: fjellrapp), \textit{Phleum alpinum} (N: fjellkjevle) and \textit{Equisetum} sp. (N: sneller) are found around most springs. \textit{Mniobryum albicans}, which is a typical cold-water moss, is easily identified by its characteristically bright green colour and its water repellent surface, where the water is concentrated in numerous separate drops. During early summer, \textit{Mniobryum} dominates and its bright green colour makes the groundwater springs clearly visible from a distance, when the vegetation elsewhere is grey and brown. Later in the summer, it occurs close to the spring outlets where the temperature is low, but a few metres downstream of the outlet, where the temperature increases, it is hidden under other plant species. Some 10 m or so away from the spring outlet it disappears completely.

The spring vegetation, especially the \textit{Epilobium} sp., is a delicacy for sheep and many other herbivorous mammals (Nordhagen 1943). Because of the great frequency of springs in the area, the aerial distribution of these plants is considerable and makes the valley particularly attractive for grazing. This is also indicated by the name ‘Godlidalen’, which means ‘the valley with rich hills’.

Discussion and conclusions

The described aquifers are related to rather coarse-grained ice-marginal deposits. Similar shallow aquifers and springs are found in inland areas all over Norway, as
well as in mountainous parts of Sweden (Resvold-Holmsen 1932; Nordhagen 1943, p. 437) and are thus a quite common hydrological feature in mountainous areas in Scandinavia. The typical characteristics of these mountain springs seems to be the same as for those described in this study, at least in areas where the bedrock is resistant to weathering and has a low content of carbonates minerals.

Individual springs may have a rather limited discharge of a few hundred litres per hour, on average. However, where marked spring horizons occur, the aggregated discharge may be considerable. The aquifers are small, with low storage capacities and transmissivities. Nevertheless, in an area where the overburden is otherwise predominantly compact basal till, with very low hydraulic conductivity, the springs comprise the major groundwater aquifers. As such, they are of great importance for the local hydrology.

The stable and significant base-flow of the River Åsta during the winter (see Englund & Haldorsen 1983) is largely dependent on the discharge from all the springs in the catchment. As shown in Fig. 2, the springs are found mainly in the eastern and northern part of the area. Tributary rivers flowing from this part of the catchment have more stable base-flows, with higher pH, cation contents, and bicarbonate values than tributaries from the western side of the catchment. In the west the bedrock is covered with a continuous and thick blanket of compact basal till, and the precipitation is discharged mainly as surface water or follows shallow channels in the soil. In the west the pH is sometimes so low that it is a problem for the fish fry (Sønsteby & Maartmann, pers. comm. 1991). This problem does not occur in the eastern tributaries, which are buffered by the water from the groundwater springs. The groundwater of the springs is thus important for maintaining the ecosystem in the catchment.

The secondary effect of the spring water discharge, that on the surrounding vegetation, was described in an earlier section. In Godlidalen, the spring areas are important grazing grounds for moose and thus also influence hunting activities. In modern times, when mountain farming has become less important, the spring water still remains an attractive water supply for cottages built close to old mountain farms or farms that are restored for recreational purposes.

The conclusion is, therefore, that the typical mountain springs, whose water has a low residence time and low ion content, constitute an important part of the hydrology in mountainous areas in Scandinavia.

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