Downward extension of the local magnitude scale for induced earthquakes in the Netherlands

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Context

After more than six decades, the Dutch government decided to close the Groningen Gas field: after hundreds of induced earthquakes, the societal license to operate had been lost. One lesson learned from this episode is the need to be able to detect and characterize induced seismicity at an early stage. Concurrently, the Dutch subsurface is more and more subject to human intervention. For example, close to 40 geothermal doublets have been drilled so far, with the majority now producing heat. These developments have spurred the government to ask the KNMI to prepare for a significant densification of its nationwide seismic monitoring network (Figure 1, left). In addition, operators are encouraged to have their local network stations included in KNMI's seismic network. As a result of this densification, the probabilistic (local) magnitude of completeness (MoC) will become lower than 1.5 for most of the country (Figure 1, right).

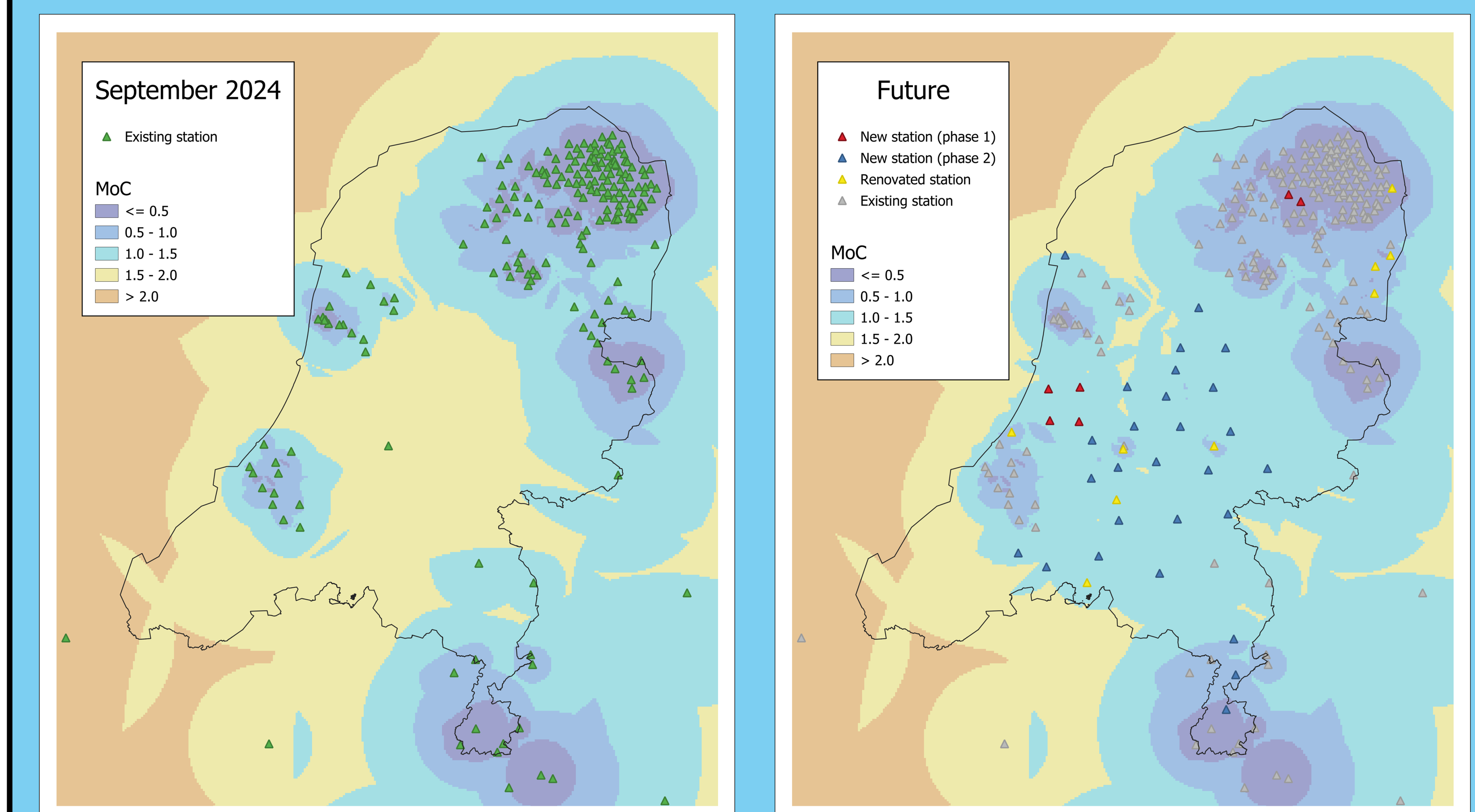

Figure 1: *MoC for the KNMI network configuration (KNMI, 1993) on September 2024 (left), and as planned in the future (right). For details regarding the MoC computation, see Ruigrok et al. (2023).*

Current procedure and local magnitude

The local magnitude M_L is derived from the peak displacement A (in millimeters) on a Wood-Anderson (WA) torsion seismometer (undamped period $T_0 = 0.8$ s, damping $h = 0.8$, gain=2800). Specifically, it is defined as (Richter, 1935)

$$M_\mathrm{L} \equiv \log_{10} A - \log_{10} A_0, \quad (1)$$

where A_0 is a correction for attenuation with distance. Richter determined the attenuation correction for California such that at an epicentral distance of 100 km, $\log_{10} A_0 = -3$.

After the installation of a first-generation network of vertical borehole strings in the nineties, the KNMI used gas-extraction induced seismic events to calibrate a local magnitude scale for such events. The calibration was performed using the horizontal components of each string's deepest geophone, typically located at a depth of 200 meters. Following Kanamori et al. (1993), this attenuation function was defined to have the functional form

$$A_0(R) \equiv cR^{-n}\mathrm{e}^{-\alpha R}, \quad (2)$$

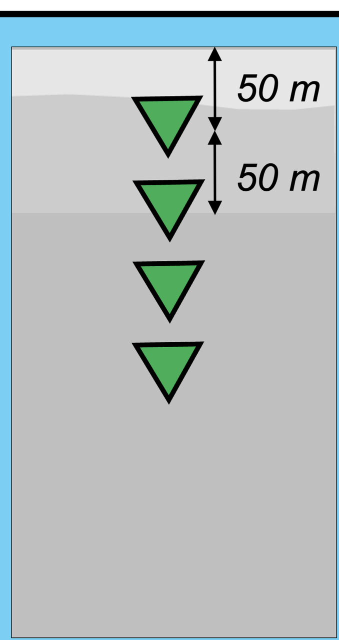

Figure 2: *Typical configuration of a KNMI borehole array.*

where R is the hypocentral distance in kilometers. Using a dataset of 157 records, values of 0.3767, 1.33, and 0.0032 were found for c, n, and α, respectively (Dost et al., 2004).

For an individual geophone, the current procedure for estimating an induced seismic event's local magnitude is as follows:

1. Retrieve the event's recording by both horizontal components (HH1 & HH2).
2. Remove the instrument response to obtain horizontal particle displacement, including a (frequency domain) pre-filter with corner frequencies of 0.125, 0.25, 50, and 100 Hz.
3. Apply a fourth order Butterworth bandpass filter (frequency band between 0.5 and 40.0 Hz).
4. Convolve the two traces with the response of a Wood-Anderson seismometer.
5. If needed, apply a Butterworth high-pass filter to suppress microseisms.
6. Retrieve the maximum absolute amplitude on the HH1 component in a ~10 second window containing the S-wave arrival.
7. Retrieve the HH2 component's maximum absolute amplitude in the same window.
8. Compute the average of these two maximum absolute amplitudes, and multiply by 1000 to obtain the WA-displacement in millimeters. This yields A.
9. Use equation (1) to obtain an estimate of the geophone-specific local magnitude.
10. Determine the Signal-to-noise ratio (SNR); discard the computed M_L if the SNR is smaller than 2.

Averaging all geophone-specific local magnitudes subsequently results in the induced event's M_L.

References

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Problem statement & Objective

For small induced events in the north and west of the Netherlands, both our current attenuation relation and procedure are not adequate. This is due to the fact the Wood-Anderson simulation promotes (relative) low frequencies, often causing the (relatively) high-frequency earthquake signal to be overwhelmed by low-frequency noise. This is exemplified in the figure to the right: the recordings of the M_L 0.43 induced event suffer significantly from the microseisms (bottom plot), whereas this is not the case for the M_L 2.07 event (top plot). The top trace represents the raw recordings and the second trace the particle velocities after deconvolution by the instrument response. The third trace is obtained after bandpass filtering (0.5-40 Hz) and convolution with the Wood-Anderson response. After application of a steep 0.9 Hz Butterworth high-pass filter (fourth trace), the microseisms still dominate the recordings of this weak event. We therefore derive an attenuation relation, which will augment the existing attenuation relation (equation (2)) in the sense that it enables a downward extension of our existing local magnitude scale.

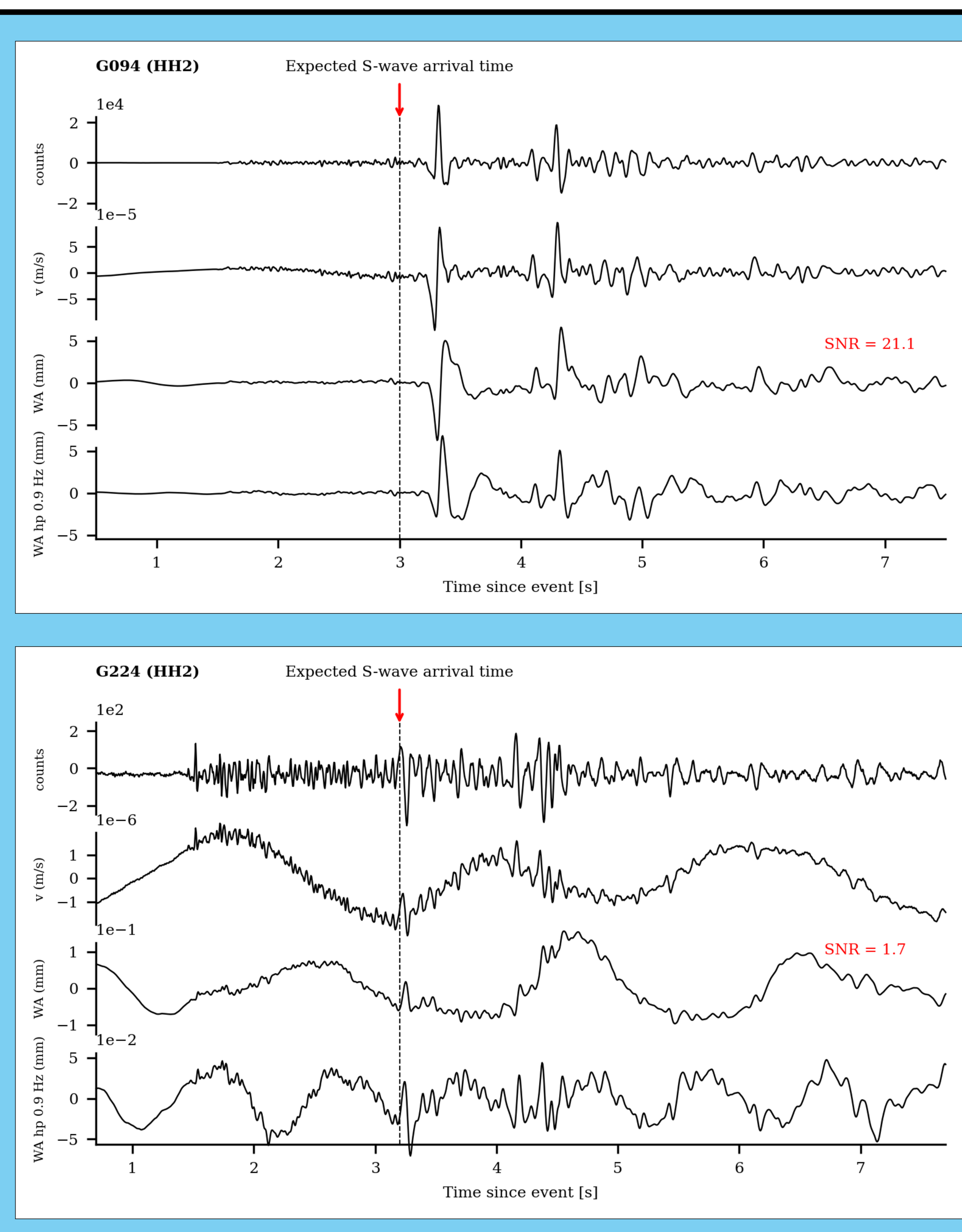

Figure 3: *Recordings by geophone G094 of a M_L 2.07 event on 2017-12-10 (top) and geophone G224 of a M_L 0.43 event on 2024-07-06 (bottom). The horizontal recordings of G094 (epicentral distance of 3.34 km) resulted in a M_L estimate of 2.09. The horizontal recordings of G224 (epicentral distance of 3.93 km) resulted in a M_L estimate of 0.31.*

An attenuation relation for small induced events ($M_\mathrm{L} < 1$)

Local magnitudes based on the newly derived attenuation relation we denote by $M_\mathrm{L}^{(v)}$. Similar to M_L, $M_\mathrm{L}^{(v)} \equiv \log_{10} A^{(v)} - \log_{10} A_0^{(v)}$, where $A^{(v)}$ is the measured horizontal peak particle velocity (in meters) and $A_0^{(v)}$ the newly derived attenuation relation. The functional form of $A_0^{(v)}$ is assumed to coincide with that for A_0 (Equation (2)). Our new attenuation relation needs to be consistent with the existing attenuation, which is enforced by minimizing

$$\varphi \equiv \sum_i^K \sum_j^{N_i} \left| M_{\mathrm{L},ij} - M_{\mathrm{L},ij}^{(v)} \right| \quad (3)$$

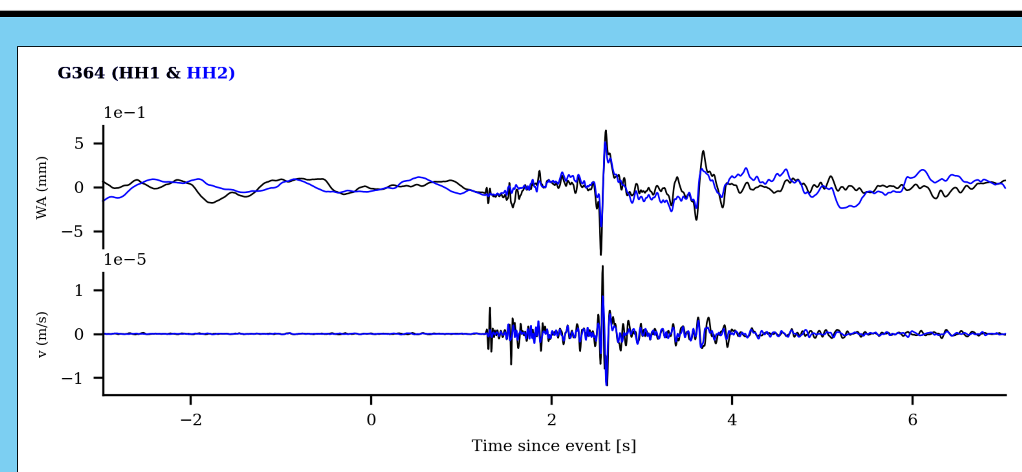

Figure 4: *Horizontal recordings by geophone G364 of a M_L 0.87 event on 2017-01-03 (epicentral distance of 1.96 km). The recordings resulted in a M_L estimate of 0.97 by this geophone for this event. Recordings by the HH1 component are depicted as solid black traces; recordings by the HH2 component as solid blue traces. The top plot represent the Wood-Anderson displacement (in mm). The bottom plot the particle velocity after application of a fourth order bandpass filter (5 - 40 Hz).*

while deriving c, n, and α governing $A_0^{(v)}$. The index i refers to the event with K being the total number of events used to calibrate the attenuation relation; the index j refers to the station with N_i being the total number of stations for which the magnitude of event i could be estimated.

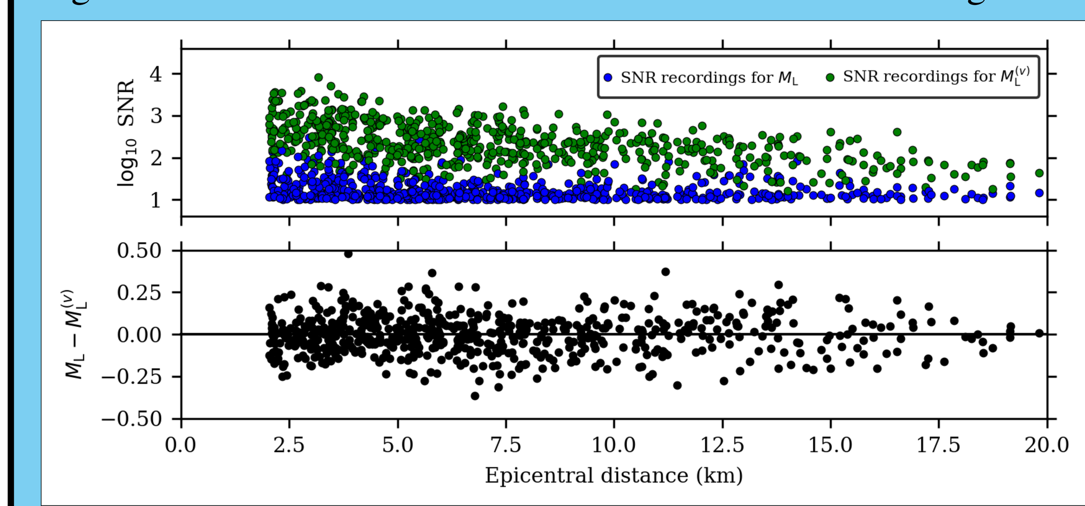

Figure 5: *Signal-to-noise ratios (top) and $M_\mathrm{L} - M_\mathrm{L}^{(v)}$ (bottom) of all recordings used to calibrate $M_\mathrm{L}^{(v)}$.*

The signal-to-noise ratios (SNRs) of the (5-40 Hz bandpass filtered) recordings of particle velocity are consistently much higher than the SNRs of the WA simulated recordings (Figure 5, top). This is in itself not surprising, and the very reason the former recordings were chosen as a basis for the augmented magnitude scale. Calibration is performed using induced events in the Groningen region in 2017 for which $0.5 < M_\mathrm{L} < 2.0$. Furthermore, only horizontal recordings for which SNR > 10 are included. Calibration based on the selected (geophone-specific) local magnitudes results in values of $9{\times}10^{-6}$, 1.38, and 0.0555 for c, n, and α, respectively.

The bottom plot of Figure 5 shows that, in most cases, the difference between the two geophone-specific local magnitudes does not exceed 0.3. We find that the difference between event-specific local magnitudes in fact does not exceed 0.15 (for events with $N_i > 3$). We therefore conclude that the derived attenuation relation is a good (first) candidate to extend the existing magnitude scale downwards.

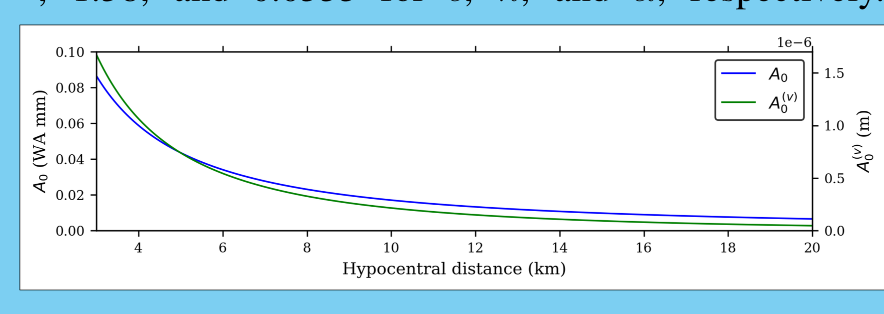

Figure 6: *Attenuation relation associated with M_L (blue) and $M_\mathrm{L}^{(v)}$ (green).*