

Mapping submarine groundwater discharge areas— an example from Laholm Bay, southwest Sweden

Abstract—We conducted a broad reconnaissance for freshwater upwelling along the south-eastern coast of Sweden from Båstad 5 km northward. A 300- × 350-m area of sandy seabed, 0–2 m deep, was selected for detailed mapping and study. Results were similar regardless of whether measurements were taken with a bottom-contact electrical-conductance sediment probe, by suction of the upper centimeter of the seabed, or with seepage meters. While the point measurements (suction samples and seepage meters) provided an estimate of seepage flux, the sediment-probe reconnaissance provided confidence that a large area of groundwater discharge was not overlooked. The nearshore discharge (0–50 m offshore) probably originates from a shallow aquifer and the offshore discharge (50–220 m offshore) from a buried, confined aquifer.

There is growing evidence that submarine groundwater discharge can be significant. Large submarine springs can often be detected visually at the sea surface (Zektzer et al. 1973). The emerging groundwater usually differs markedly from the surrounding seawater and may create a plume, which can be measured with physical or chemical parameters (Hay 1984; Baker et al. 1987; Hovland and Judd 1988; Rosenberg et al. 1988). Smaller submarine springs and seeps and the dispersed seepage through the coasts and sea floor are usually less spectacular and more difficult to study, yet often represent the principal form of submarine groundwater discharge (Zektzer et al. 1973). Several methods for locating and estimating groundwater discharge are available (Zektzer et al. 1973; Lee 1977, 1985; Johannes and Hearn 1985; McLachlan 1989). Three of them are further developed and compared here. It is suggested that they may form a complementary set.

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During the last decade, serious fish kills and oxygen depletion in near-bottom waters have been reported from various south Swedish and Danish coastal areas. Nitrogen losses from extensively cultivated coastal lowlands, which cause algal blooms in coastal waters, are believed to be responsible for most of these problems (Fleischer et al. 1987). Coastal lowlands drain toward the sea via rivers but also via direct groundwater runoff. Submarine groundwater discharge represents only a minor portion of the total freshwater input. It may, however, contribute significantly to the eutrophication and pollution of coastal waters because the concentrations of NO_3^- and other solutes in groundwater are frequently higher than the concentrations in other freshwater inputs and several orders of magnitude higher than the concentrations in unpolluted coastal waters.

Vanek and Gustafsson (1988) investigated nutrient inputs by groundwater to Laholm Bay. They found that groundwater discharge in the bay was usually concentrated within a 10–30-m-wide nearshore strip and decreased rapidly farther offshore—a pattern well known from other freshwater and saline environments (Lee 1977; Vanek 1987). Moreover, denitrification within the organic-rich nearshore sediments seemed to remove most of the groundwater-transported NO_3^- before it reached the sea (Vanek and Gustafsson 1988).

Methods used in the preliminary study included compilation of geological and hydrological data, reconnaissance drillings, installation of observation wells, thermal infrared scanning, geoelectrical and seismic soundings, pore-water salinity surveys along the shoreline, and seepage meter measurements (Vanek and Gustafsson 1988). In one area an extensive freshwater seepage zone of unknown origin was observed some 50–80 m offshore. It was difficult to describe

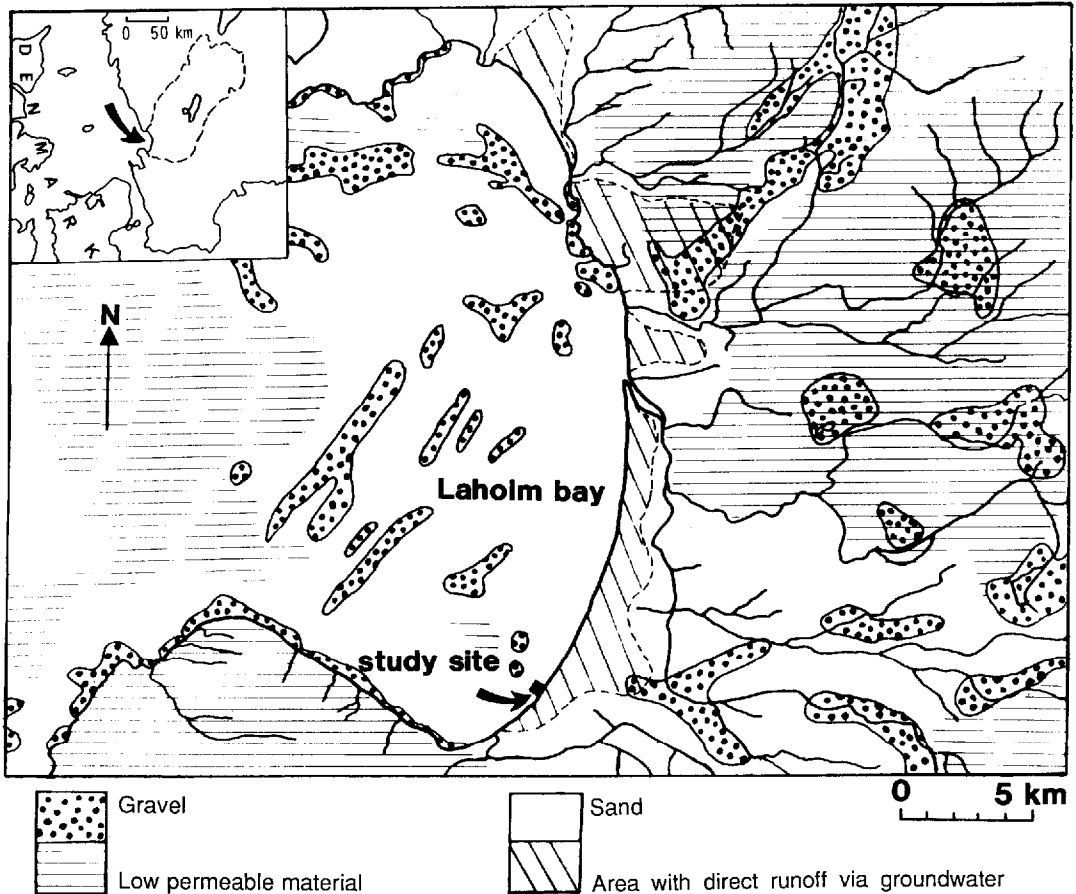


Fig. 1. Location of study site and surficial geology of Laholm Bay and the adjacent coastal plain. Insert: Location of the bay and its drainage area (dashed line) in southwestern Sweden.

groundwater flux in this zone with the above methods, and we began a detailed investigation to test other techniques better suited for mapping and sampling submarine seepage zones.

Laholm Bay is on the southwestern coast of Sweden in a shallow part of the Kattegat. Water salinity above the halocline (at a mean depth of 14.5 m) usually ranges between 12 and 22‰ (corresponding to ~18–31 mS cm⁻¹ at 20°C). Under the halocline it is higher, typically between 30 and 33‰ (42 and 45 mS cm⁻¹ at 20°C). Bottom sediments are sandy over 65% of the area. At depths >10–15 m, however, the sandy layer is often thin and underlain by glacial clay (Floderus and Håkanson 1989; Fig. 1).

The land near the bay is covered by a 2–7-m-thick layer of eolian, highly permeable

sand. This layer creates a shallow, unconfined aquifer. Mean groundwater discharge from this aquifer to the sea was estimated to be ~15–20 liters km⁻¹ of shoreline per second (Vanek and Gustafsson 1988). The surficial sandy layer is underlain by a thick sequence of low-permeable glacial clay. Several glaciofluvial eskers (long, winding ridges of sand and gravel) buried under the clay create a lower, confined or semiconfined aquifer. There is no information on the hydraulic connection between these deeper aquifers and the sea, though some investigators (e.g. Mörmner 1969) have suggested connections may exist. The region is characterized by small tides (~20 cm), frequently overridden by barometric pressure effects. During autumn storms, sea level may rise 150 cm or more.

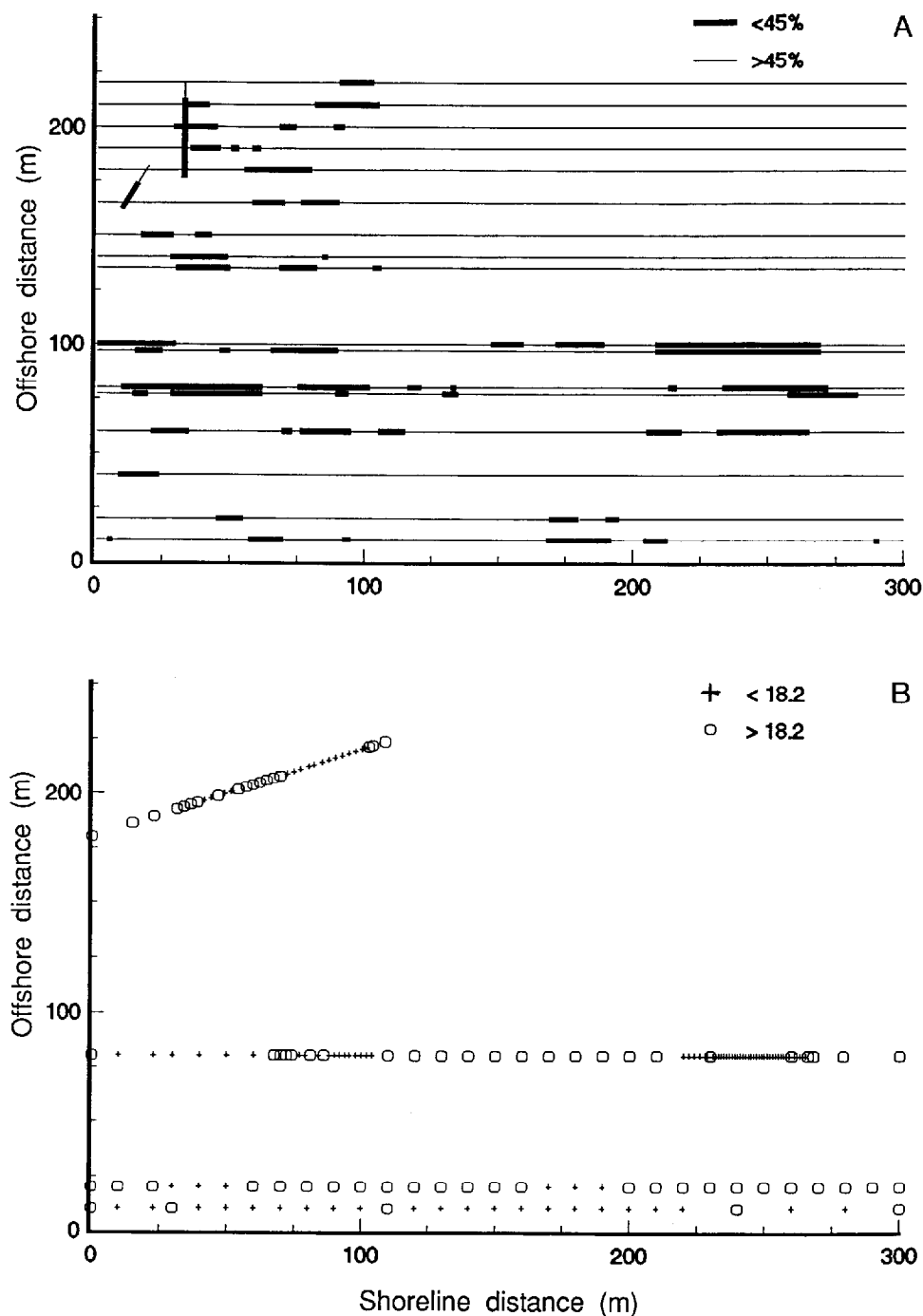


Fig. 2. Conductivity probe profiles (A) and shallow pore-water conductivities (B) in a 300- × 300-m nearshore area near Skummeslöv. Freshwater seepage zones are those with bulk sediment conductivity <45% of local seawater (panel A, bold lines) or shallow pore-water conductivity <18.2 mS cm⁻¹ (panel B, +). Zero horizontal line corresponds to the mean high-tide shoreline, and X-Y coordinates are those used in Figs. 4–7 and 9 and in the text.

Our study site was a limited area just south of the Skummeslöv beach in the southern part of the bay (Fig. 1). The seabed here is covered by fine-rippled sand ($d_{50} = 0.13$ mm) which is often redeposited during storms. The sand layer is of variable thickness (typically between 0.1 and 2 m) and is usually underlain by low permeability glacial clay. In addition, large stones and minor gravel areas (occasionally buried in the sand) have also been observed and seem to be associated with observed freshwater seeps. Initial surveys confirmed that most freshwater seepage is concentrated within a 300×220 -m grid area, where reference points consisting of wooden sticks and floats were laid to facilitate the determination of *X-Y* coordinates (Fig. 2).

From a preliminary study (Vanek and Gustafsson 1988), the location and extent of freshwater seeps near Skummeslöv indicated a deeper, but unknown source of groundwater discharge, and local geological conditions appeared rather complex. To investigate these seepage zones, we used the following methods. We dragged a continuously recording probe behind a boat to measure the bulk electrical conductivity of the surficial sediments as described by Lee (1985), except that a recording chart was also included. We used a suction sampler to sample pore water and measure the shallow pore-water electrical conductivity in the surficial sediment layer. The sampler consisted of a porous filter tip, a protective plastic plate, a flow-through chamber containing a conductivity-measuring probe, and a supporting hard plastic rod (Fig. 3). When in use, the plastic plate with the filter tip was pressed slightly on the sediment surface, and pore water from the uppermost centimeters of the sediment was drawn to

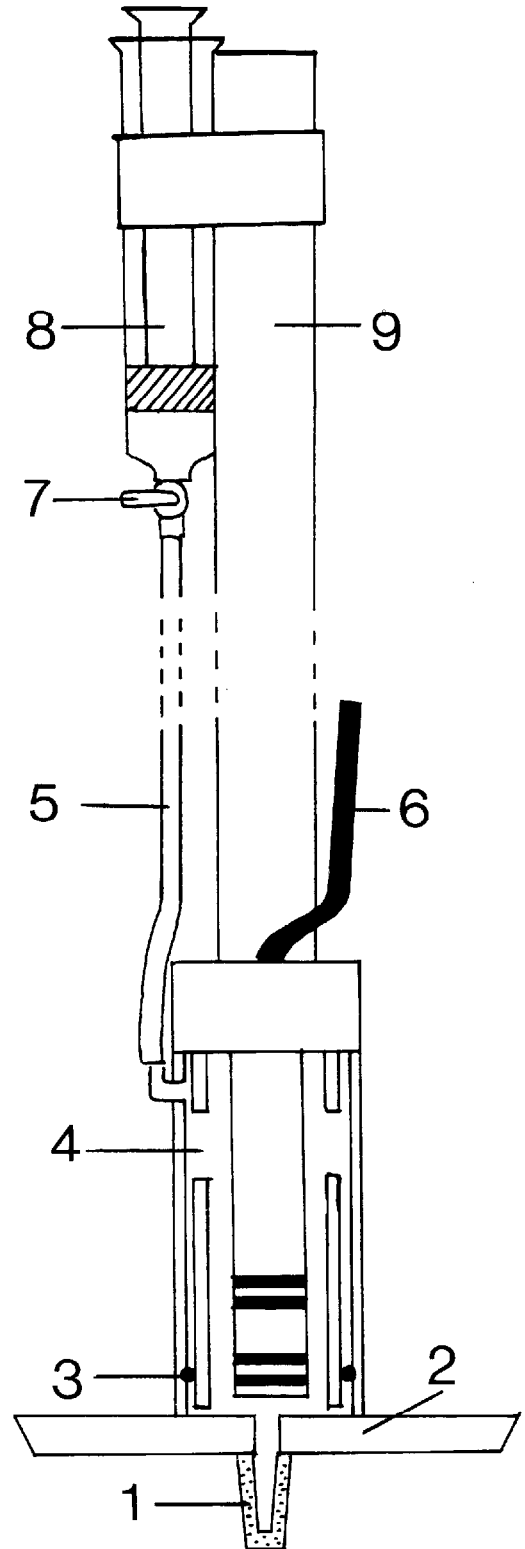


Fig. 3. Pore-water conductivity measuring probe for shallow, sandy sediments. 1—Filter tip of sintered bronze, porosity $40 \mu\text{m}$ (Krebsöge GmbH); 2—hard PVC plate, 10-cm diameter; 3—O-ring; 4—flow-through chamber with standard conductivity cell (Hanna Instr.); 5—PVC tubing, 3-mm i.d.; 6—connecting cable to a field electrical conductivity meter; 7—3-way valve; 8—50-ml plastic syringe; 9—support rod of hard PVC, 150 cm long.

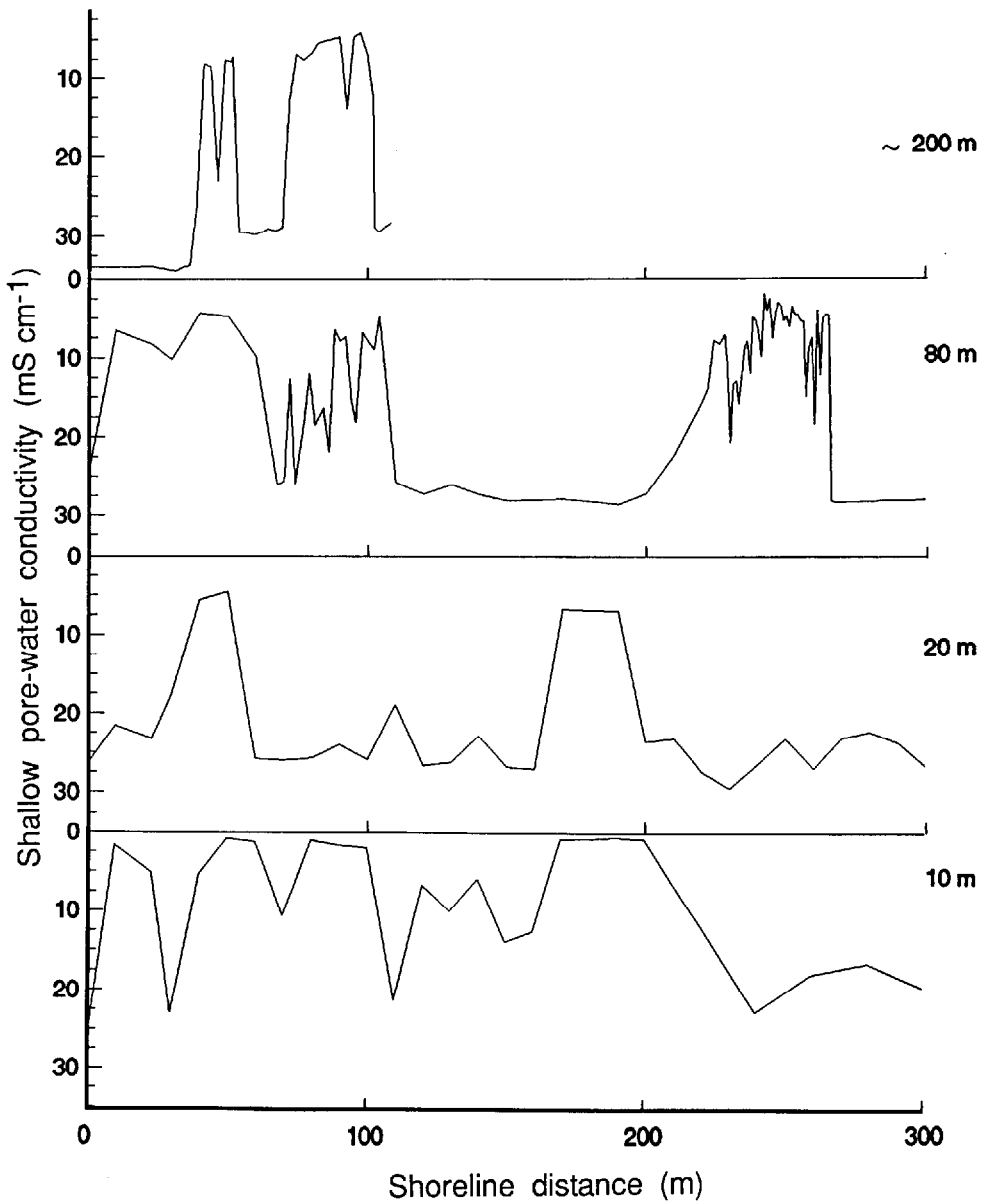


Fig. 4. Shallow pore-water electrical conductivity transects running parallel to the shore, at Y distances of ~200 (180–220), 80, 20, and 10 m.

the flow-through chamber by applying suction on the connected plastic tubing. Pore-water conductivity was read on a meter display as soon as the value stabilized. From 10 to 30 ml of water were needed to flush the filter tip and the flow-through chamber.

We also measured bulk electrical conductivity of the surficial sediment by push-

ing a standard laboratory conductivity measuring probe directly into the sediment to a depth of 1–3 cm. Values so obtained are partly influenced by tortuosity and other sediment characteristics (Berner 1980). In sandy sediments, however, the differences in bulk conductivity due to low-salinity anomalies are usually significantly higher

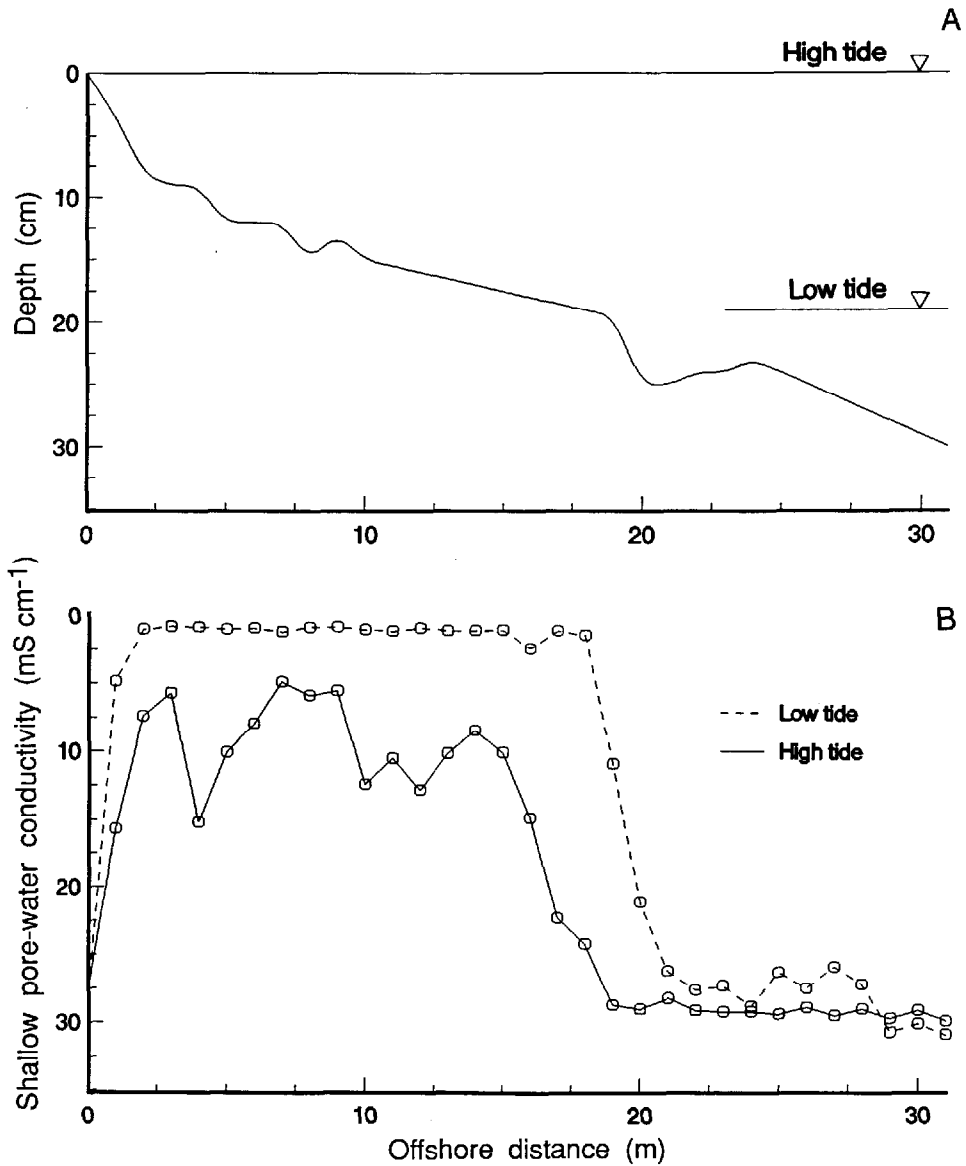


Fig. 5. Water depth (A) and shallow pore-water electrical conductivity (B) profiles at the transect perpendicular to the shore (X distance, 190 m; Y distance, 0–31 m) during high and low tide.

than the variability caused by the sediments themselves (Vanek and Gustafsson 1988).

We used seepage meters similar to the original design of Lee (1977), but smaller (0.0625 m²) and made of plastic barrels (25 liters). Collecting bags of light polyethylene, with 5-liter maximal volume, were connected to the meter via 12-mm-i.d. silicone tubing. It was determined later (Shaw and

Prepas 1989; Vanek in prep.) that unprotected bags of this kind used with zero initial water volume tend to draw in water and result in erroneously high seepage estimates. To correct for bag-associated errors, we subtracted the mean seepage rate for places with groundwater flow assumed to be zero. It is probable that the bag-associated errors diminish with higher seepage rates and may

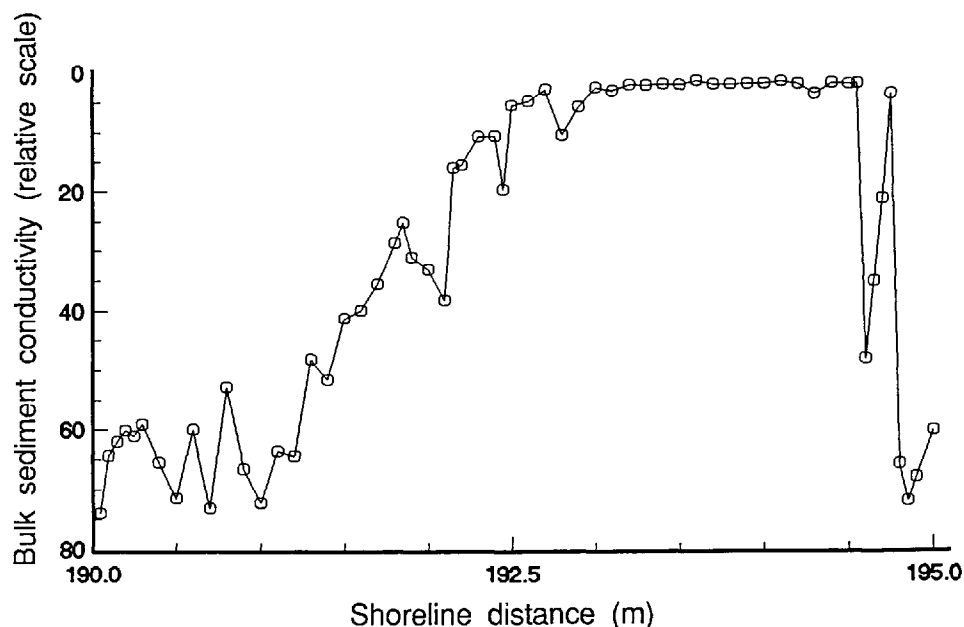


Fig. 6. Bulk sediment electrical conductivity along a transect parallel with the shoreline (X distances, 190 and 195 m; Y distance, 20 m).

be low when using meters directly in highly active seepage zones.

We used minipiezometers for water sampling that were similar to those described by Vanek and Gustafsson (1988). There were no attempts to measure vertical hydraulic gradients, except for several qualitative observations of positive gradients, which indicated upward potential for flow through the sediment within active seepage zones.

Nutrients and major ions in water samples were analyzed with a Technicon AutoAnalyzer, and tritium with a Packard Auto-Gamma spectrometer. Water samples for nutrient analyses were preserved by adding HgCl_2 ; all samples were stored in the dark and cold.

Most measurements were done in June 1988. After the initial tests, over 20 line-kilometers of bulk sediment conductivity were obtained by the dragged conductivity probe. Most of the measurements were inside the 300- $\times$ 350-m grid in transects parallel with the shoreline. Low-conductivity sediments, suggesting possible freshwater influence, were found in all profiles located between 10 and 220 m offshore (Fig. 2A). Outside this area, no offshore low-conduc-

tivity anomalies were found, except for a minor area outside the harbor of Båstad, ~ 5 km southwest of Skummeslöv. Shallow pore-water conductivity was measured along four profiles parallel to the shoreline at 10, 20, 80, and ~ 200 m (180–220 m) offshore (Figs. 2B and 4). Mean weighted value of pore-water conductivity in the grid was 21.05 mS cm^{-1} , whereas mean seawater conductivity was 28.2 mS cm^{-1} and ground-water conductivity in the nearest water-supply well (230 m inland, ~ 20 m deep) was 0.6 mS cm^{-1} .

In three small nearshore areas, the distribution of low-salinity anomalies was studied in more detail. During both high and low tide, 32 pore-water conductivity measurements were made every meter along a transect perpendicular to the shoreline at an X distance of 190 m. During low tide and minimal wave action, lower values of salinity were observed than during high tide (Fig. 5).

In the same area during high tide and 20 m offshore, bulk sediment conductivity was measured with a standard laboratory measuring probe at 60 points spaced every 5–10 cm between X distances of 190 and 195

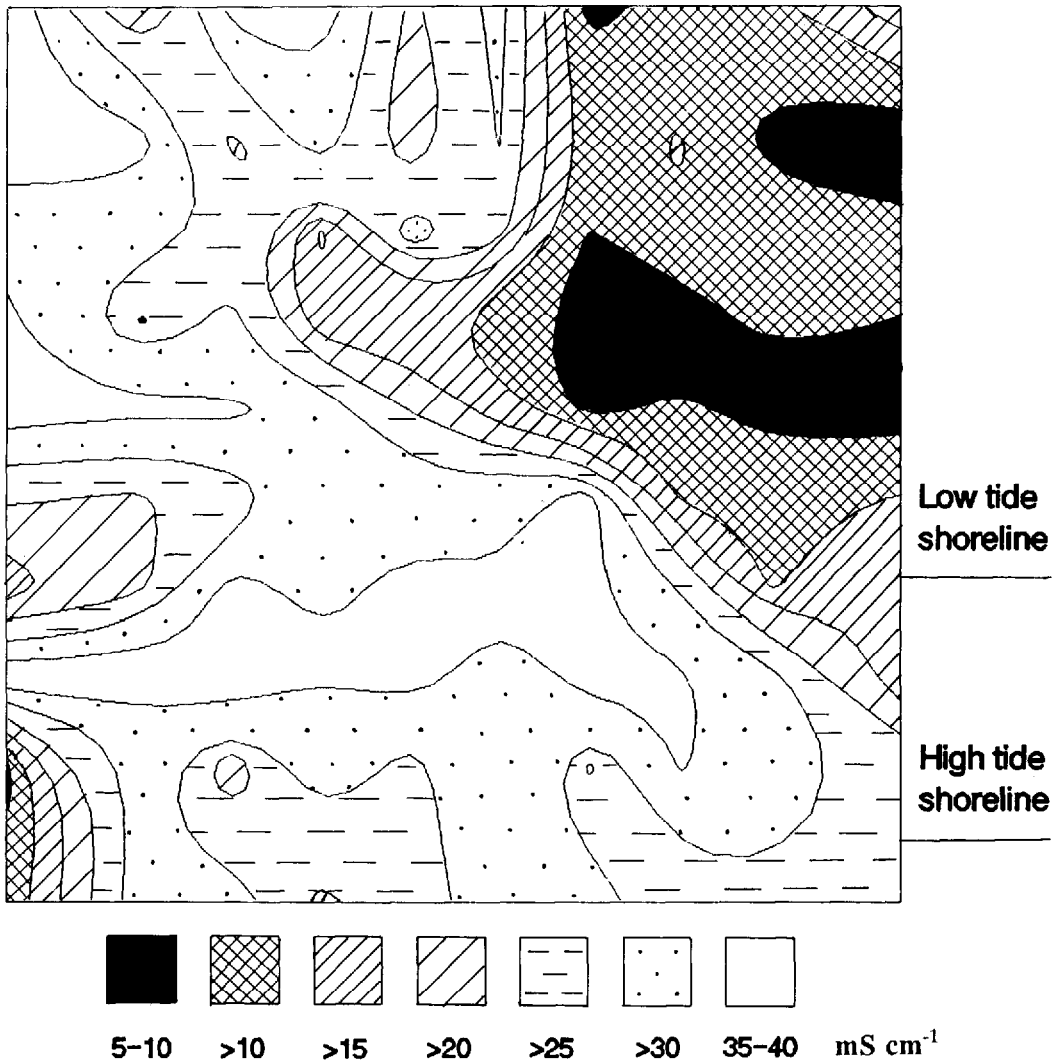


Fig. 7. Distribution of low electrical conductivity anomalies within a 20- × 20-m square area, based on 420 regularly spaced point measurements of shallow pore-water conductivity. Picture was created with GRIDZO software (X distances, 40–60 m; Y distances, -1 to + 19 m).

m. Low bulk conductivities due to fresh-water influence were observed in the area between 192.0 and 194.5 m near a large stone (Fig. 6).

In November 1988, surficial pore-water conductivity was sampled in a 20- × 20-m nearshore area located between X distances of 40 and 60 m (grid spacing, 1 m). Mean conductivity of pore water in the area was 24.7 mS cm^{-1} , which corresponds to a mixing of 61.5% seawater (39.8 mS cm^{-1} at this time) and 38.5% groundwater (0.6 mS cm^{-1}).

In spite of waves 0.5–1 m high that brought in new seawater and mixed it with shallow pore water, several isolated low-conductivity anomalies were easily observed (Fig. 7).

In 1987–1988, seepage meter velocity measurements in low-salinity areas (as defined in Fig. 2) ranged between 30 and 150 m yr^{-1} , ($x = 90 \pm 9 \text{ m yr}^{-1}$, $n = 23$). Outside these areas, where groundwater influence was low or negligible, seepage rates of 3–24 m yr^{-1} were recorded ($x = 16 \pm 1 \text{ m yr}^{-1}$, $n = 26$).

Table 1. Groundwater and seawater chemistry near Skummeslöv site, June 1988. 1—20-m-deep well; 2—seepage meter 10 m offshore; 3—minipiezometer 80 m offshore; 4—minipiezometer 200 m offshore; 5—seawater; 6—typical range from shallow wells in the region. All values in mg liter⁻¹, unless stated otherwise; values from points 1–5 based on only one sample.

	1	2	3	4	5	6
Sp cond. (mS m ⁻¹)	56.5	76.8	28.0	70.8	2,600	20–30
pH	7.77	8.26	7.80	7.89	—	—
Alk. (meq liter ⁻¹)	2.65	3.44	3.42	3.97	1.30	—
Ca	47.4	32.4	61.2	60.1	270	—
Mg	6.55	22.4	52.0	12.1	680	—
Na	59.4	93.8	476	68.2	—	—
K	4.21	8.48	22.4	8.88	—	—
Fe	0.06	0.06	0.33	1.29	—	—
Mn	0.07	0.06	0.11	0.14	—	—
(NO ₃ ⁻ + NO ₂ ⁻)-N	0.246	0.002	0.002	0.001	0.150	2–10
NH ₄ ⁺ -N	0.012	0.350	0.162	0.345	—	—
Cl	96.1	137	847	125	10,000	—
SO ₄ ²⁻	26.1	41.5	139	11.0	1,450	—
Tritium (TU units)	26 ± 9	138 ± 13	–9 ± 9	–13 ± 7	33 ± 15	—

To identify the origin of the discharging groundwater, we took water samples from the nearest well located 230 m inland, from a well-flushed seepage meter located 10 m offshore, and from two minipiezometers located 80 and 200 m offshore and 40 cm and 80 cm below the seabed, respectively. Water chemistry data were then compared with the water chemistry of seawater and shallow groundwater from nearby wells (Table 1).

In the area of Skummeslöv there are two aquifers from which groundwater may originate: a shallow, unconfined aquifer outcropping near the shoreline which probably delivers 2–3 liters s⁻¹ of groundwater along the 300-m-long shoreline section, and a deeper, confined aquifer, probably connected with an esker which outcrops 3 km inland near the village of Skottorp (Karlqvist et al. 1985). It is reasonable to expect that tritium concentrations are highest in the shallow groundwater, intermediate in seawater, and low or zero in the deeper aquifer, which in 1988 was discharging water with tritium content lower than the background concentrations found before atmospheric testing of nuclear weapons. By analyzing the data in Table 1, the following conclusions can be drawn. The well 20 m deep on the shore contains a mixture of shallow groundwater with low conductivity and high NO₃⁻ and tritium concentrations, and deep groundwater characterized by higher conductivity and low NO₃⁻ and tri-

tium concentrations. Groundwater discharging 10 m offshore is tritium-rich, with minimal influence by seawater, and originates from the shallow, unconfined aquifer. Groundwater discharging 80 and 200 m offshore contains no tritium and low Na and Cl concentrations and must originate exclusively from the deeper aquifer.

Laholm Bay, as well as other marine coastal areas receiving river inputs, is characterized by unpredictably varying salinity. This may complicate mapping groundwater discharge areas because pore-water salinity may not be in equilibrium with near-bottom seawater. The time needed to reach 99% equilibrium between salt water and the sediment was calculated with Ogata's (1970) diffusion-advection model for a nonreactive solute. From Fig. 8 it follows that 99% equilibrium should be achieved within a day for shallow (<3 cm) sediment depths and advection rates ≥3 m yr⁻¹.

Groundwater flux is only one advection process that causes water movements within the sediments. Other processes include tide and wave-induced currents, bioturbation, thermal and density-driven advection, sedimentation, and sediment compaction rate. There is only limited information on the magnitude of these processes in different kinds of sediments and environments. It is, however, unlikely that the advection rate in marine coastal sediments, even those not affected by groundwater flow, would be <3

m yr^{-1} . According to Riedl et al. (1972), wind and wave-induced currents cause advection streams of $\sim 3.5 \text{ m yr}^{-1}$ in shallow-water (0–200 m) sediments. Of the same order of magnitude or higher are swash-driven filtration rates (Riedl and Machan 1972). According to McLachlan (1989), the typical range of advection velocities for different beach types and various tide and wave regimes is $0.1\text{--}70 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1}$ of shoreline, which for the 20-m-wide nearshore strip corresponds to advection velocities of $1.8\text{--}1,300 \text{ m yr}^{-1}$. Thus, advection rates of 3 m yr^{-1} or higher are probably encountered in most nearshore sandy sediments. This allows one to assume that the solute chemistry of sediment pore water can be controlled largely by groundwater discharge even along moderately high-energy beaches and to use the conductivity ratio between shallow pore water and near-bottom seawater as a measure of freshwater influence.

Pore-water salinity patterns within periodically exposed intertidal sediments are often complex. In addition to groundwater and seawater effects, pore-water salinity may be influenced by direct precipitation, evaporation, and dissolution of previously deposited salt. Low pore-water salinities associated with the low-tide period may also be caused by surficial freshwater runoff originating in a visible seepage zone further inland, as in Fig. 5.

There is reasonably good agreement between the measurements of bulk sediment conductivity and surficial pore-water conductivity (Fig. 9). In addition, there is a positive correlation between shallow pore-water conductivity and seepage rate (Fig. 10). Thus, the estimation of total groundwater flux can be made with continuous records of bulk sediment conductivity for interpolating between points of known pore-water salinity or seepage rate. However, further modifications of the dragged probe will be needed to lower the scatter of background data which may be considerable.

The data in Fig. 2 indicate that the total area of active seepage zones near Skummeslöv is between 12,000 and 15,000 m^2 . Taken together with the mean (corrected) seepage rate of 75 m yr^{-1} , the total flux of fresh groundwater is between 28 and 35 li-

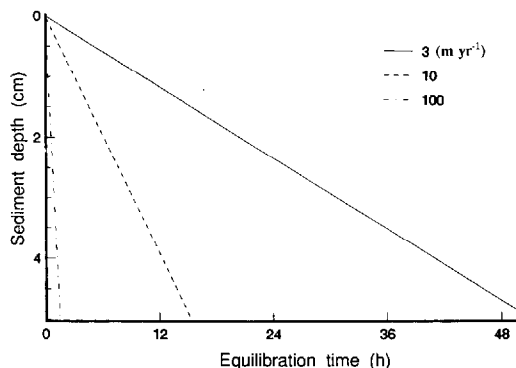


Fig. 8. Time for 99% equilibration of chloride concentration between sediment and overlying water for three different advection velocities, calculated according to Ogata (1970) with the following assumptions: seawater salinity, 16‰; sediment salinity at time zero, 0‰; effective diffusion coefficient, $D = D^* n^2$, where D^* is the self-diffusion coefficient for chloride ($1.6 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$) and n is sediment porosity (35%). Diffusion and advection are considered to act in opposite directions.

ters s^{-1} . A similar estimate (35 liters s^{-1}) was obtained from the mean weighted pore-water conductivity value for the $300 \times 220\text{-m}$ grid (21.05 mS cm^{-1}) and the relationship between pore-water conductivity and seepage rate (Fig. 10). Both estimates, however, are largely dependent on seepage meter measurements, which were crudely corrected for the the bag-associated errors. To obtain correct seepage meter measurements, one should use prefilled bags in combination with several exposure times and subsequent correction for anomalously high initial water fluxes into the bags (Shaw and Prepas 1989).

Theoretically, submarine groundwater discharge zones can be mapped with only point data such as pore-water conductivity or seepage meter measurements. However, our results on the patchiness of seepage zones (Figs. 2, 6, and 7) show that many points will be necessary to localize the borders of previously unknown, often highly irregular, anomalies. In some cases such as localization of submarine springs in karstic regions, thermal infrared scanning and similar remote-sensing techniques may be useful (Gandino and Tonelli 1983; Roxburgh 1985). In our case, however, the thermal anomaly on the sea surface, caused by the

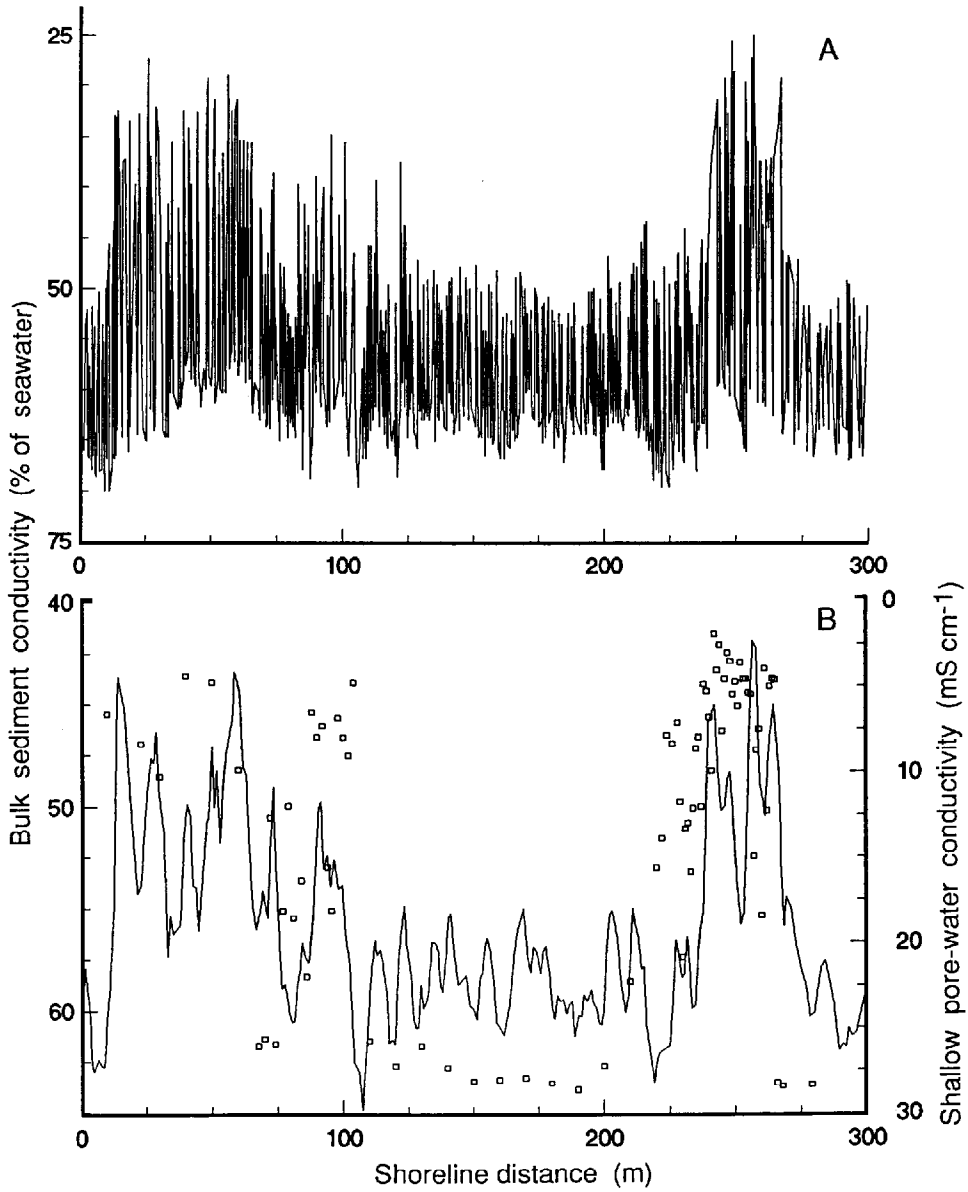


Fig. 9. Comparison of dragged probe record (A—continuous record; B—solid line, running 5-m means) and shallow pore-water conductivities (B—□) at the transect parallel to shore, Y distance, 80 m.

inflowing groundwater, was probably too small to be recorded, or it may have been obscured by overlying seawater (Gustafsson et al. 1988). Geophysical measurements and reconnaissance drillings on shore as well as pore-water salinity and temperature measurements near the shoreline have proven too insensitive for detailed investigations of

discharge areas farther offshore (Vanek and Gustafsson 1988). Our data suggest that the use of a dragged sediment probe in combination with point measurements of shallow pore-water conductivity and seepage rates are likely to give better results.

In our experience, the three methods described form a complementary set. The sed-

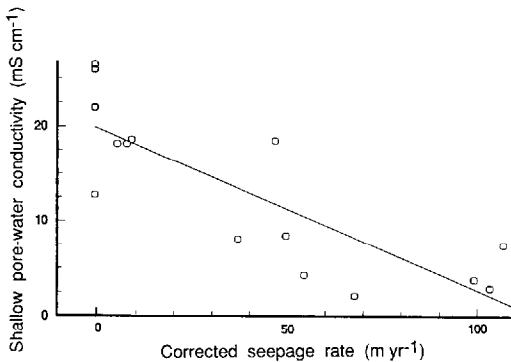


Fig. 10. Linear correlation between shallow pore-water electrical conductivity (y) and corrected seepage meter measurements (x) in low-conductivity sediment zones near Skummeslöv ($y = 19.98 - 0.175x$, $r^2 = 0.67$, $n = 15$, $P \leq 0.01$).

iment probe can be used to cover large areas fairly rapidly (1 m s^{-1}). Bulk electrical conductivity of the sediment may vary with sediment type and is not necessarily groundwater-specific. Possible freshwater influence, therefore, must be confirmed by some other method, such as sampling of shallow pore water or measurement of seepage.

The shallow pore-water sampler is much slower, ~ 100 point samples per hour, but provides a value that can be directly compared with standard laboratory determinations. In contrast to the dragged sediment probe, the sampler is easy to make and can also be used to obtain pore-water samples. While recognizing that such samples may represent a mixture of seawater and pore water, potentially they can be used for other solute analyses. In its present form, the sampler is limited to sandy bottoms and shallow water depths.

Seepage meters are more labor intensive, providing only one measurement per hour of effort on average. When used properly, seepage meters allow direct and accurate estimates of seepage rates within a broad range of values. The value obtained is integrated over the area of the meter, which should result in more representative values than point measurements obtained with conductivity probes or piezometers and may be particularly important in nonhomogeneous sediments with patchy seepage zones. Seepage meters are easy to construct and operate,

particularly in shallow depths. Seepage meters can also be used for water sampling, provided that the groundwater flow is strong enough to flush the meter efficiently.

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