4 Hydrological Measurement

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4.1 INTRODUCTION

The previous policy-oriented chapter identified a number of actions that hydrologists can take to ensure that hydrological science contributes to policy making and resolution of the current and anticipated world water resource issues. Hydrological measurement underpins many of these action points and is the focus of this chapter. The current state of the art in the measurement of parameters and processes that fall within the hydrological expertise possessed by the Working Group members is provided in this chapter. Some of the gaps and shortcomings are discussed and the chapter concludes with an examination of future prospects and development in hydrological measurement.

Our understanding and ability to model hydrological processes are often impeded by issues of *data* and *scale*. The characterization (that is, how we measure a specific quantity) of a hydrological process is a key element in hydrological activities. Cuts in monitoring programmes, particularly in precipitation networks, have created a general shortage of data yet there is an increased need for hydrological data and greater accuracy in spatial information. Addressing the issue of scale is essential and particularly important as measurement scales are space and time dependent.

This chapter recognizes the importance of measurement in advancing our understanding of hydrological science and for improving our ability to model, predict and manage. Hydrological measurement requires immediate attention if significant progress is expected to be made since we currently lack the tools to effectively monitor and therefore understand hydrological science. However, there are several obstacles to progress: hydrologists are not making their instrumentation needs known to those involved in instrument design; the hydrological community in general lacks funds to develop specific hydrological instruments; instrument design and data collection have a diminished status from the perspective of funding agencies and researchers; and technological development is not considered a science. These prejudices need to be removed in order for hydrological scientists to increase their capacity for measurement.

In very general terms, quantitative hydrology is concerned with the estimation of one or more of the terms within the hydrological water balance, at a range of spatial and temporal scales. However, even in the well known and inherently simple mass balance expression for Q, the net flux out of a system (whether that is extraction or discharge) (equation 4.1):

$$Q = P - E - \partial S / \partial t \tag{4.1}$$

where, P is rainfall, E is evapotranspiration and $\partial S / \partial t$ is the change in surface storage per time step, many practical difficulties arise in the measurement of each term. Most hydrological measurements are point measurements, or, at best, integrated over a variable, but typically unknown, area or volume.

Hydrological systems are classically divided into surface and subsurface systems that comprise the storage and movement of water at and under the surface of the earth. Surface water includes all water bodies that are in direct contact with the atmosphere (i.e. streams, lakes, snow and ice, as well as water in the biosphere). Subsurface water includes the traditional definition of "groundwater" (water that occurs in the permanently saturated (or phreatic) zone beneath the water table) and water in the aerated unsaturated (or vadose) zone above the water table. The global reserves of water are shown in Table 4.1. Freshwater resources are mainly stored in glaciers and permanent snow cover as well as groundwater. The proportion of water stored in the atmosphere, soil, and in river channels is very small, and the dynamic residence times are short. Nevertheless this transient water plays a crucial role in the global hydrological cycle and understanding of the related processes is the fundamental aim of hydrological research. The water fluxes within the global hydrological cycle are shown in Fig. 4.1. Since they are extensively discussed in all classical hydrology textbooks (e.g. Dingman, 2002) they will not be examined in detail here.

As Table 4.1 and Fig. 4.1 indicate, mean residence times of water in different compartments of the hydrological cycle vary greatly from 2500 years to as little as nine days, with flux rates also being highly variable. The status of hydrological storages and, subsequently, the generated runoff are subject to large spatial and temporal variabilities and heterogenous distributions. Knowing residence times, total stored volumes and flux rates is key to our ability to accurately predict and model the distribution and movement of water and associated solutes and sediment. The accurate

Table 4.1 *World water reserves (from Korzun, 1978 and Oki et al., 2004); residence times are calculated as the mean volume divided by the mean flux.*

Form of water	Covering area (km^2)	Total volume (km^3)	Mean depth (m)	Share of volume (%)	Mean residence time
World oceans	361 300 000	1 338 000 000	3700	96.539	2500 years
Glaciers and permanent snow cover	16 227 500	24 064 100	1463	1.736	56 years
Groundwater	134 800 000	23 400 000	174	1.688	8 years
Ground ice in zones of perma-frost strata	21 000 000	300 000	14	0.0216	
Water in lakes	2 058 700	176 400	85.7	0.0127	
Soil moisture	82 000 000	16 500	0.2	0.0012	
Atmospheric water	510 000 000	12 900	0.025	0.0009	9 days
Marsh water	2 682 600	11 470	4.28	0.0008	
Water in rivers	148 800 000	2 120	0.014	0.0002	18 days
Biological water	510 000 000	1 120	0.002	0.0001	
TOTAL WATER RESERVES	510 000 000	1 385 984 610	2718	100.00	

[a] Excluding Antarctic groundwater (approximately 2 000 000 km^3).

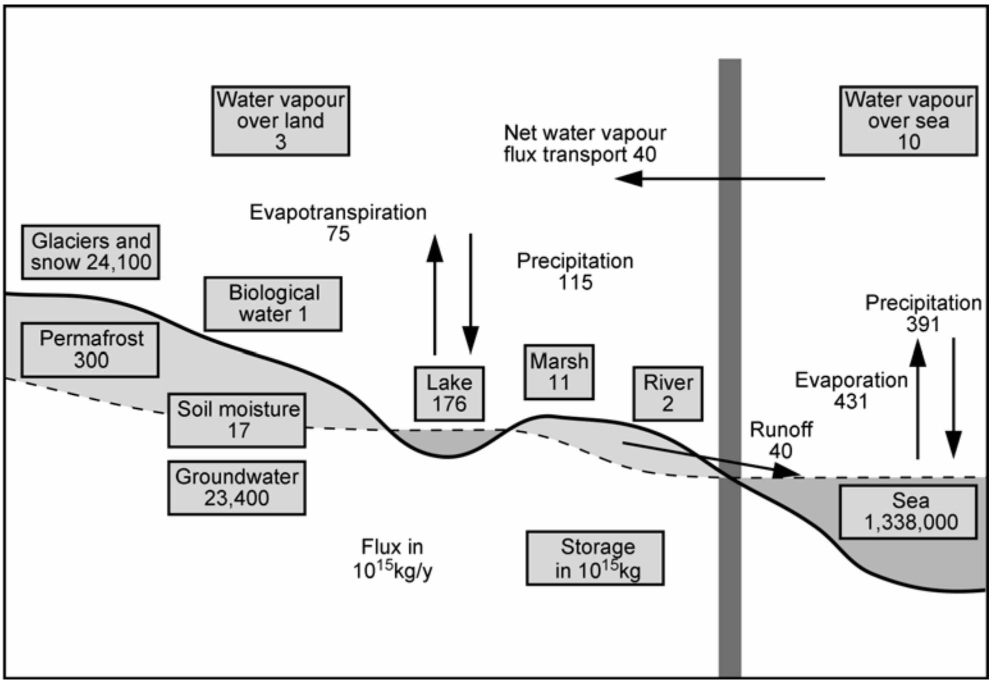

Fig. 4.1 *Schematic of the global water cycle. Water storages (in 10^{15} kg) from Table 4.1 are shown in the boxes and approximate fluxes (in 10^{15} kg year^{-1}) for the period 1989–1992 are represented by arrows (after Oki et al., 2004).*

measurement of these quantities is essential for managing and developing appropriate water resources policies. Currently, the *quantification* of hydrological fluxes at different spatial and temporal scales remains insufficiently understood and greater effort must be made to measure these fluxes.

There is a large range of scales governing what hydrologists are required to measure, making accurate quantification a daunting task. Data are required both at short (hours or days) and very long time scales (decades) so maintaining long-term databases is vital to understanding hydrological water balance. Furthermore, the relevant scale of measurement for each term in the water balance equation above can vary depending on the application. For example, urban hydrologists may require precipitation intensities every minute while global water balance hydrologists only require average monthly precipitation. Similarily, the spatial scale of measurement may vary from sub-metre scales to hundreds of kilometres. Therefore, appropriate water resource modelling and maangement requires the following ingredients: an excellent understanding of the scale of the problem, data that respect the scale, and an appreciation of the errors in the data.

4.2 PRECIPITATION

Precipitation is the primary input to all hydrological models and its accurate quantification is one of the key sources of error in modelling and prediction.

Precipitation in general is one of the most highly variable climatic parameters and varies across a range of space–time scales (New *et al.*, 2001). Although advances in monitoring precipitation have been made across a variety of scales, numerous studies demonstrate that error induced by inaccurate precipitation measurement is a major hurdle in hydrological characterization.

Measurement of rainfall has traditionally been achieved by direct sampling through ground-based networks of gauges. While a variety of gauge designs exist, they all represent essentially point measurements. Ground-based radar and satellite-based sensors can provide spatially continuous estimates of rainfall and have become essential to characterizing precipitation in the last few decades, but the raingauge will probably always be an integral part of measurement networks.

4.2.1 Ground-based measurement techniques

The raingauge Gauge data is generally collected over land but coverage is non-uniform, particularly at high latitudes, in arid regions, and in parts of the tropics (New *et al.*, 2001). Although it is the only measurement method available in many regions of the world and for short-term high intensity events that are important in urban hydrology (New *et al.*, 2001), gauge monitoring networks are declining primarily due to economic or political pressures.

There are numerous errors associated with raingauge measurements, including: observer errors; undercatch caused mainly by wind-induced turbulence; splashing; and evaporation. Correction methods are documented in Seibert & Morén (1999) and are known to work well for longer time intervals. Correction for individual events is far from trivial requiring many variables that may not be measured at the gauge site. Tipping bucket gauges are by far the most widely used device for rainfall measurement (Habib *et al.*, 2001), but Ciach (2003) showed that local random errors are substantial in the measurement of short high-intensity rainfall events by tipping bucket gauges. A possible solution is to design networks in which stations have two or more gauges since using two independent measures of the same quantity can reduce local random errors (Ciach, 2003; Hanson e*t al.*, 2004). Within the NOPEX project (NOrthern hemisphere climate-Processes land-surface EXperiment) Seibert & Morén (1999) tested a raingauge with a new type of windshield (INSITU IS200W) that reduced both wind-induced and wetting losses for rainfall measurements.

Ground-based radar Radar measurements of rainfall are completely different from raingauge measurements both with regard to the scale of measurement and the actual quantity being measured. In contrast to the point, one-dimensional measurements made by raingauges, radar can provide three-dimensional structural information on storms over large distances. There are several methods of rainfall estimation by radar but techniques for accurate estimation with high spatial and temporal resolution are still unsatisfactory as a large number of meteorological and hardware factors influence the quality of the measurements (Sauvageot, 1994). Improvements both in measurement and processing conditions are possible, but the main shortcomings for radar estimates arise from *target homogeneity*. Consequently, Sauvageot (1994) suggests that it is better to limit the maximum distance of radar observations and develop a network of radar systems, each surveying a restricted area, rather than having fewer stations observing over large distances. The global distribution of ground radar stations is, unfortunately, similar to that of raingauges, in that, due to the costs involved, they tend to be found in more prosperous regions and in areas of high population density (Kidd, 2001). Orbital radars on satellite platforms offer very attractive and promising responses to global rainfall monitoring (Sauvageot, 1994).

4.2.2 Remote sensing of precipitation

Remote sensing has been used for decades to provide estimates of precipitation. Most algorithms dealing in the visible, near infrared and thermal infrared wavelengths utilize information on cloud height and coldness which are both related to the amount of precipitation that is expected from the cloud. These methods are suitable for monthly rainfall totals and have been recommended for estimating climatic scale distributions of convective rainfall often found in the tropics and subtropics during warm seasons. The GOES (Geostationary Operational Environmental Satellite) precipitation index, for example, is one of the simplest and most widely used infrared indices of precipitation in the tropics and subtropics. In addition, there are numerous algorithms that rely on empirically derived functions of passive microwave measurements from sensors such as SSM/I (Special Sensor Microwave Imager). Microwave methods do not represent surface precipitation rates but are actually vertical integrations of rain and/or ice water content in the atmosphere. Poor temporal sampling, due to deployment of microwave sensors on polar satellites, reduces the precision with which short-term precipitation can be estimated. Passive microwave techniques tend to provide more accurate estimates of instantaneous rain areas and rates as compared to infrared techniques, while infrared algorithms tend to outperform passive microwave techniques over longer time periods. Algorithms that involve calibration with raingauges generally provide better rainfall estimates than satellite-only techniques but performance tends to be location dependent. Statistical testing of the validity of different techniques has found merged products to be greatly superior to all other products, but comparison is problematic since the validation data tend not to be independent of the merged product (Kidd, 2001).

Multi-sensor algorithms attempt to combine the benefits of frequent infrared imagery with microwave methods, but the calibration and validation of satellite-based rainfall estimation algorithms are unresolved problems in hydrology. These algorithms can only be evaluated successfully if there is an appropriate reference standard of known accuracy. Because raingauges provide point measurements with often questionable accuracy, sparse raingauge networks do not provide sufficient information for realistic representations of rainfall fields and are thus of limited use for calibrating and validating satellite-based algorithms for precipitation. Improvement in the use of remotely-sensed estimates of precipitation requires quantifying the sources of errors in the *in situ* data being used to develop the algorithm (Kidd, 2001).

A series of intercomparison projects have been undertaken to determine the strengths and weaknesses of satellite rainfall algorithms and to help develop better algorithms for meteorological and climatological products. For example, the Precipitation Intercomparison Projects showed that, while there is variability in precipitation estimates at instantaneous scales, the algorithms produced good estimates in the regions and meteorological conditions for which they were developed or calibrated and multi-satellite approaches appeared to work well. One problem, common to all intercomparison exercises, was the limited quality and amount of validation data which ultimately limited the usefulness of the results (Kidd, 2001).

The current state of the art in satellite remote sensing of rainfall is exemplified by the Tropical Rainfall Measuring Mission (TRMM), launched in 1997, in which data obtained at various wavelengths are combined and merged. The TRMM makes measurements in the visible, near infrared, thermal infrared, and passive microwave wavelengths and has an active radar system to enhance our understanding of rainfall (Kummerow *et al.*, 1998). Although remotely-sensed estimates of precipitation have been used operationally as input to hydrological models in a number of places such as the Sudan and Egypt, most operational satellite applications for precipitation use visible or infrared wavelengths. Satellite-based precipitation products are currently available through, or sponsored by, a variety of agencies, including NOAA, NASA, WMO and NCDC (National Climatic Data Center).

As the future of estimating precipitation using remote sensing may rely on merging satellite data with raingauge data, global precipitation databases will increasingly be required. The development of such databases has encompassed initiatives to collate historical precipitation gauge measurements, to increase the representation and reduce errors in existing public domain gauge data, and to develop techniques for retrieving remotely-sensed data (New *et al.*, 2001). Global precipitation datasets include those held within the Global Precipitation Climatology Centre (GPCC), part of the World Climate Research Programme; the Global Historical Climatology Network (GHCN); and the Global Summary of the Day (GSD).

4.3 INTERCEPTION AND EVAPOTRANSPIRATION

Most hydrological models are comprised of “loss” and “routing” parts. Routing is modelled fairly well with the current knowledge of hydraulics, but the loss part is still an elusive quantity to both measure and model for a variety of reasons. “Loss” is considered to be the intercepted water or water stored in depressions, soil or biota that is eventually transferred from the land surface to the atmosphere by the processes of evaporation and transpiration.

4.3.1 Interception

Losses due to interception are primarily a function of the *interception storage capacity* of the vegetation and the *evaporation opportunity* (Eagleson, 1970). The interception storage capacity of the vegetation can be related to the characteristics of the canopy cover and is often modelled using Leaf Area Index (LAI) or some other canopy characteristic. The evaporation opportunity is a function of the nature of the precipitation event and can vary from negligible quantities to values exceeding the interception storage capacity of the canopy. Eagleson (1970) stated that observations of interception loss were highly variable and “have not yet yielded to generalization”. This statement still remains true today and, consequently, interception in models is often a parameter calibrated to match observations of net runoff.

Horton (1919) conducted one of the first detailed studies to quantify the role of interception and proposed an equation for modelling interception. Since then numerous models of interception have been formulated, ranging from single parameter deductions from gross precipitation to slightly more sophisticated models such as the Gash model (Gash, 1979) or the more data intensive Rutter model (Rutter *et al.*, 1975) used in models such as MIKE-SHE (Refsgaard & Storm, 1995). Various modified versions of these models are continually being developed and tested today (for example, Carlyle-Moses & Price (1999) found the error in the estimation of interception using a revised Gash model amounted to less than 1% for a northern hardwood stand).

The effects of precipitation type (snow *vs* rain), canopy nature (deciduous *vs* coniferous and broadleaf *vs* needles (Huber & Iroume, 2001)) and canopy structure (the significance of stemflow, understorey vegetation and multiple layered canopies (Price *et al.*, 1997)) on interception have been studied more closely in recent years. Huber & Iroume (2001) reported that the assumption that interception is negligible where there are multiple canopy layers may only be valid for dense canopies which do not allow incident radiation to penetrate sufficiently to evaporate intercepted precipitation from secondary layers of the canopy. Hedstrom & Pomeroy (1998) observed that intercepted snow quantities are related to branch strength and stand age (which is no doubt also applicable to rainfall interception).

Regardless of these distinctions, both the interception storage capacity of the vegetation and the evaporation opportunity are routinely modelled as functions of LAI. LAI can be estimated over small scales with *in situ* optical based measurement devices such as the LAI-2000 plant canopy analyzer produced by LI-COR, Inc. in the USA, or the TRAC (Tracing Radiation and Architecture of Canopies) instrument made in Canada. LAI is also easily determined over large scales through the use of remote sensing (see below). Intuitively LAI would be thought to exhibit significant correlation with interception and this initial supposition was confirmed by Pierce & Running (1988) who showed that interception was directly proportional to LAI. However, examination of the theoretical relationship between interception and LAI concluded that LAI and canopy storage capacity exhibited a linear relationship only in areas with low to medium LAI (Van Dijk & Bruijnzeel, 2001). The only likely condition in which LAI and canopy evaporation would exhibit a linear relationship, according to the Penman-Monteith model, would be when an increase in LAI resulted in a considerable reduction in aerodynamic resistance of the canopy. Alavi *et al.* (2001) concluded that interception losses can be estimated with reasonable accuracy on a seasonal time scale for dense spruce stands using only LAI and aerodynamic resistance as input variables.

4.3.2 Depression storage

Losses to runoff generation may also occur from water stored in surface depressions that is eventually evaporated. Depressional storage capacity (DSC) has often been estimated from its positive correlation with surface roughness (Guzha, 2004). Rougher surfaces have a greater DSC and hence DSC is often estimated from a random roughness index (RR, normally computed as the standard deviation of height measurements) (Hansen *et al.*, 1999). Infiltration, evaporation and runoff retardation are all influenced by random roughness.

A wide variety of parameters and indices have been developed for estimating surface roughness and these have been tested against measurements or estimates of DSC. Hansen *et al.* (1999) noted that a quantity called “mean upslope depression” had the highest correlations with DSC, but, if the RR index was used to estimate DSC, then heights would require adjustments related to slope orientation. From their study, Kamphorst *et al.* (2000) reported that RR was the best choice for computing DSC from all the indices tested. However, a predictive model based on RR would contain an error of 3 mm because RR does not consider the spatial configuration of the surface which is essential for computing depressional storage.

The indices mentioned above are essentially physically-based mathematical models of DSC developed for use in a number of applications, including agriculture and quantification of erosion. One of the main problems in their verification is the arduous process of determining the true DSC of a surface. Microrelief meters can be used to measure surface roughness but are laborious to use, even over small surfaces. A much more efficient method which is becoming more common and inexpensive is the use of terrestrial laser scanners, which can generate a data cloud of millions of elevation measurements for areas of tens of square metres. With a number of scans, a complete hillslope can be represented so that it can be easily visualized, manipulated and analysed to determine a wide array of useful parameters for hillslope hydrology.

4.3.3 Evapotranspiration

Evapotranspiration is perhaps the most difficult hydrological flux to measure or model, especially at regional scales and greater. Major progress in linking evapotranspiration processes with energy exchange came about in the 20th century where the work of Penman (1948) laid the foundation for relating evapotranspiration to meteorological variables (ASCE, 1998). Today, one of the most commonly used methods for estimating evapotranspiration is the “Penman-Monteith” method.

Evaporation from water surfaces (often a key component in water balance studies) is rarely measured directly, except over relatively small spatial and temporal scales, but is instead computed from evaporation pans, water balance, energy balance, mass transfer procedures and combined methods. The choice of one technique over another is largely a function of the data availability and the required accuracy of the computed evaporation (ASCE, 1998). Evaporation pans provide estimates of potential evaporation but they cannot represent vegetative controls on moisture loss. Additionally, the use of pans is complicated by feedbacks with the moisture content of the overlying near-surface atmosphere; if the actual evaporation rate is low, then the atmospheric demand for moisture is high, thus elevating pan losses, and *vice versa*. To account for these feedbacks, empirical “pan coefficients” are employed as corrections.

In the majority of water resources applications, the key water loss estimate required is evapotranspiration from land surfaces, particularly when soil water is limiting. When soil water is not limiting, evapotranspiration can be readily estimated using the reference crop coefficients or the Penman-Monteith equation. However, physically-based models like the Penman-Monteith equation often require large amounts of detailed information that may be impractical to obtain over large areas (Jiang *et al.*, 2004).

A wide variety of instrumentation and approaches are available to measure evapotranspiration, including gravimetric methods, neutron scatter methods, and measurements of soil water potential. The latter method can be inaccurate due to a number of factors related to variability within the site, calibration needs, and the sampling procedure that may alter the characteristics of the site (ASCE, 1998). All these methods suffer from scaling issues; that is, how to extrapolate from a point measurement to larger areas.

Making direct measurements of evapotranspiration irrespective of soil water conditions might be considered ideal; however, the challenge is to extrapolate point or localized measurements to large spatial scales or other time periods. Direct evapotranspiration measurements are difficult, time-consuming, expensive, require upkeep/maintenance, and have certain fetch requirements (ASCE, 1998). Recently, the Bowen ratio and eddy correlation methods have been devised to estimate actual evapotranspiration losses from land surface sensible and latent heat fluxes. These fluxes are not directly measurable quantities, but are usually calculated from other more directly measurable parameters based on well-established theories or empirical relationships (Jiang *et al.*, 2004). While both methods represent a significant improvement on the use of pans, fluxes are determined from an area (or fetch) of between 100 and 1000 m^3 that varies according to wind direction and strength and the convective stability of the atmospheric boundary layer. While areally-integrated measures of evapotranspiration are provided, it is clear that significant variability of evapotranspiration may exist within the area of the fetch. From many studies, Jiang *et al.* (2004) reported that the Bowen ratio method has an uncertainty of about ±10% or 10 W m^{-2} (whichever is larger) in estimates of surface sensible and latent heat fluxes, while for the eddy correlation method at least ±10% uncertainty exists for surface fluxes. For a relatively simple formulation, the lower bound for uncertainty in estimating surface sensible heat flux may depend on the accuracy of the basic physical quantities being measured, such as surface and air temperature, wind speed, and roughness length (Jiang *et al.*, 2004). For large heterogeneous areas, or under unstable atmospheric conditions, the uncertainty is expected to increase. Thus it may be argued that more complicated approaches do not necessarily lead to greater accuracy in evapotranspiration estimates, unless uncertainties in input data and model parameters can be controlled.

Improving technology to decrease errors in evapotranspiration estimates should be a goal of future research. Remote sensing can also play a role in scaling up point estimates to regional estimates by providing insight into the spatial and temporal variability of several key parameters and variables. For example, the estimation of evapotranspiration requires information on the amount of available radiative energy, the type/nature of the surface and the presence of water that can be evaporated.

Remote sensing for monitoring evapotranspiration The use of remote sensing for estimating evapotranspiration has been investigated for more than two decades. Because of the number of variables affecting evapotranspiration, remote sensing has been used as input at a wide variety of levels by exploiting all possible ranges of the electromagnetic spectrum that are available.

Whether evapotranspiration is estimated using a simple equation based solely on surface temperatures (Caselles *et al.*, 1998) or from a SVAT model (Olioso *et al.*, 1999), remote sensing at a variety of wavelengths has been used to characterize surface temperature, soil moisture, structure of vegetation canopies or LAI, albedo, water vapour, maximum canopy conductance, photosynthetically active radiation and photosynthesis for an unstressed canopy, as inputs to evapotranspiration models. Engman (1999) cites numerous examples of successful applications of remote sensing, such as using thermal remote sensing for estimating regional evapotranspiration (Price, 1982), deriving net radiation in the Priestley-Taylor model (Kotoda *et al.*, 1983) and using satellite-derived surface temperature data in conjunction with wind and temperature soundings (Brutsaert *et al.*, 1993) to give good estimates of the sensible heat flux over forested areas on a regional scale.

The most common wavelength exploited by remote sensing applications for inputs to evapotranspiration models is in the thermal range. Surface temperature estimates in the thermal range are routinely used (Kite *et al.*, 2001) in equations of regional evapotranspiration and the split-window technique (Becker & Li, 1990) is commonly used to correct for atmospheric contamination. Thermal infrared measurements have been used to estimate sensible heat flux from the surface and other heat balance components, including latent heat flux (if the total incoming radiation is known). Accurate estimates of evapotranspiration have been obtained using only satellite derived temperature and no surface measurements. For example, Granger (1995) used a feedback relationship between surface temperature and vapour pressure deficit to estimate areal evapotranspiration that compared favourably with eddy correlation estimates from a number of surface sites (Pietroniro & Leconte, 2000). Remotely-sensed radiometric surface temperature is often used as a surrogate for actual surface temperature. This is appropriate for bare soil surfaces but not for most naturally vegetated surfaces where radiometric and actual surface temperatures are often not equal. Land surface temperature mapping from space remains a challenging problem because of atmospheric contamination, variations in geometry among sun, sensor, and target, and estimation uncertainties related to land surface emissivity (Jiang *et al.*, 2004). Similar problems affect the use of visible and near infrared wavelength ranges for the purpose of estimating evapotranspiration.

Wavelengths in the visible or near infrared range are often used to determine either land cover type, LAI (Liu *et al*., 2003) or the Normalized Difference Vegetation Index (NDVI) which are in turn used in models of evapotranspiration. The latter quantity in particular has proven useful as a number of NDVI-AET (Actual Evapotranspiration) relationships have been successfully established (Kondoh & Higuchi, 2001; Szilagyi, 2002). However, the significant relationship between NDVI and radiometric temperature (T_{rad}) assumes a complete canopy cover which does not always exist, and nonlinear relationships have been shown between the slope of the $NDVI\text{-}T_{rad}$ curve and the canopy resistance. Another approach to characterizing the role of vegetation in evapotranspiration defines the ratio of soil heat flux to net radiation in terms of vegetation cover which in turn is related to optical wavelength reflectances.

Microwave applications may seem to hold the most promise for remote sensing of soil moisture, a necessary element in estimating evapotranspiration. However, the use of passive microwave data may be restricted to regional and global scales due to resolution. Active microwave has higher resolution, but backscatter is strongly affected by soil roughness, topography and canopy architecture, making it more difficult to extract surface soil moisture. Few attempts have been made to use microwave-derived surface temperature for computing evapotranspiration (Kustas & Norman, 1996).

From measurement and modelling studies and theoretical considerations a series of issues has emerged that significantly affect the ability of remote sensing to provide estimates of evapotranspiration:

– Coincident measurements of air temperature with radiometric measurement are not typically available.
– The dependence of radiometric temperature on view angle continues to be neglected.

– Heterogeneity in thermal emissivity, atmospheric corrections and satellite calibrations contribute significant errors in the measurement of thermal radiation.
– Remote observations are instantaneous, while integrated fluxes are desired on hourly, daily or longer time scales.
– Temporal or spatial resolution and atmospheric (including cloud) contamination all affect the accuracy of estimates.

The importance of accurately characterizing the canopy Leaf Area Index (LAI) is defined as the ratio of the foliage area to the ground area within a given area. It has been determined to be important for models of evapotranspiration and photosynthesis because it is a measure of the surface available for radiation absorption and material transfer (Kergoat, 1998). Variations within the LAI can be indicative of changing moisture conditions and can result in a shift in the water balance due to differences in evaporation and transpiration. The amount of evaporation is dependent on the LAI in terms of the amount of precipitation intercepted and the amount of radiation able to penetrate the canopy. As shown in the Penman-Monteith equation, the amount of transpiration is a function of stomatal resistance in the canopy which in turn depends on the LAI (Baldocchi *et al.*, 1987; Zhang *et al.*, 2001). Transpiration exhibits a curvilinear relationship with LAI (Kergoat, 1998). Overall evapotranspiration maintains a direct relationship with LAI because when additional foliage occurs within the system more moisture is required to maintain vegetation health. As canopy LAI increases, the level of drought intensifies correspondingly.

In situ sampling of LAI is spatially limited and costly so researchers have tried to relate remote sensing measurements to *in situ* LAI measurements. Remote sensing models have traditionally attempted to use spectral vegetation indices like NDVI to model variations in LAI. However, these models have achieved only moderate success because their accuracy is often dependent on the influence of the understorey and background effects below the canopy. A review of the state of the art in estimating LAI with remote sensing (McAllister, 2005) noted two promising techniques for remote estimation of LAI: linear spectral mixture analysis and modification of spectral vegetation indices. These techniques offer explicit strategies for the mitigation of background effects but they exhibit varying results in different land covers. McAllister (2005) developed the “normalized distance method” which uses the middle-infrared range of the spectrum and provides excellent estimation of LAI from remote sensing in both needle and broadleaf canopies. Through spatial statistical analyses, McAllister (2005) reported that the quality of landscape-level LAI estimates depends on the number of pixels under consideration, the quality of the remote estimation model and the landscape variability in terms of LAI. Furthermore, the multi-scale analysis demonstrated that the remote estimation techniques exhibited significant dependence on scale and imaging platform, almost irrespective of the remote estimation model used.

4.4 SURFACE MOISTURE, RUNOFF AND DISCHARGE

4.4.1 Infiltration and soil moisture

Once moisture losses through interception, depression storage and evaporation are accounted for, moisture is expected to either infiltrate into the soil regenerating subsurface stores, or perhaps to accumulate on the surface producing surface runoff. The amount of infiltrated water can lead to an estimate of soil moisture which can be used to estimate surface runoff or water available for evapotranspiration. Conversely, if the soil moisture is known, the amount of infiltrated water, and therefore, the potential for subsurface recharge, can be estimated. Soil moisture is also considered an internal state variable that can assist in the calibration of rainfall–runoff models. Regardless of the perspective, infiltration, and the resulting soil moisture are considered by many to be the crux of the runoff generating process, and therefore at the heart of water resources modelling and management.

4.4.2 Ground-based soil moisture measurements

Both point sensors and more spatially extensive geophysical approaches have been used to measure soil moisture content. Soil moisture content is straightforward to measure for single points using the destructive thermogravimetric method, but this technique is laborious and prohibits measurements of changes in soil moisture content over time and space, which is often of interest. Methods have been developed since the 1960s for *in situ* assessment of volumetric soil moisture content by measurement of other soil properties that can be related to soil moisture content. For example, one of the first techniques developed, the neutron probe, relates soil moisture content to neutron scattering whilst more recently-developed methods use time domain reflectrometry (TDR) in which soil moisture content is related to the soil dielectric constant. The basis of these methods and their advantages and disadvantages are reviewed in Topp (2003).

To measure soil moisture content in the field with these methods a single instrument can be moved between field sites or sensors can be embedded in the appropriate soil horizon/profile to make continuous measurements. Potential uncertainties in the application of all methods for *in situ* assessment of soil moisture content include: (1) how representative the soil moisture measurement is of field conditions, since installation of the sensor/probe may perturb soil hydrology; (2) the actual volume of soil that the measurement represents; (3) the effect of soil electrical conductivity; and (4) the appropriateness of the manufacturer’s standard calibration for the measured soil. A comparison of *in situ* measurements of soil moisture content by Walker *et al.* (2004) found systematic differences of up to 10% between different sensors.

Existing methods can only provide point measurements of soil moisture content but assessments over larger spatial areas are required for inputs to landscape-scale hydrological models and for compatibility with other spatial datasets, e.g. land use and vegetation data and digital elevation models. A major breakthrough in hydrological science would be the ability to reliably assess soil moisture content beyond the soil surface over large spatial areas by geophysical methods or by remote sensing. Geophysical methods, such as tomography and ground penetrating radar (GPR), are beginning to be used for non-invasive assessment of soil moisture and flow in the unsaturated zone. GPR methods use electromagnetic energy at frequencies of 10 to 1000 MHz to probe the subsurface. At these frequencies, and in electrically resistive environments, the subsurface dielectric constant distribution governs the GPR responses. Analogous with TDR techniques, information about the dielectric constant of the shallow subsurface can be obtained from velocity analysis of GPR data via analysis of different components of the recorded signal, including analysis of transmitted energy recorded between boreholes and of direct and reflected energy acquired using surface-based acquisition geometries. Petrophysical relationships, such as the widely used Topp *et al.* (1980) empirical relationship, can be employed to link dielectric constant values obtained from GPR velocity data to estimates of water content. Experiments conducted by several investigators using different GPR acquisition geometries and analysis approaches have illustrated the value of GPR data for providing accurate and high resolution estimates of shallow soil moisture content in a minimally (or non-) invasive manner (e.g. Hubbard *et al.*, 1997; Binley *et al.*, 2002; Grote *et al.*, 2003; Lunt *et al.*, 2005). A review of the use of different GPR techniques for soil moisture estimation is provided by Huisman *et al.* (2003).

Validation and inter-comparison campaigns that collect soil moisture data can provide opportunities for modelling and algorithm development. Some of these include NASA's collaborative Land Data Assimilation Systems (LDAS), the HAPEX-Sahel campaign in 1992, the Washita experiment conducted in 1992 and the Southern Great Plains (SGP) experiments undertaken in 1997 and 1999. The USA has several operational soil moisture measurement networks, including the USDA-NRCS Soil Climate Analysis Network (SCAN); the DOE Atmospheric Radiation Measurement (ARM) – Cloud and Radiation Testbed (CART) facility in Oklahoma and Kansas; the Illinois State Water Survey soil moisture network; the Oklahoma Mesonet; and the High Plains Regional Climate Center network in Nebraska (Moran *et al.*, 2004). Currently there is no coordinated global soil moisture network, but the Global Soil Moisture Data Bank at Rutgers University, New Jersey, USA, is gathering and distributing worldwide soil moisture data. The design of *in situ* networks to validate satellite-based estimates of soil moisture is problematic because many authors have shown that soil moisture varies over multiple scales (Moran *et al.*, 2004).

4.4.3 Remote sensing for estimating soil moisture

Much effort has been expended to use remote sensing to provide regional scale estimates of soil moisture. Few applications use visible or near infrared wavelengths for determining surface soil moisture due simply to the fact that optical remote sensing measures the reflectance or emittance from only the top few millimetres of the surface at most. Predictions are confounded by the fact that soil reflectance is also affected by soil type and observation conditions (Jensen, 2000). Thermal wavelengths, either alone or in combination with vegetation indices, have been correlated with variations in soil moisture but this is not a generally widespread approach. The most promising approaches developed thus far involve the use of passive and active microwave remote sensing.

Passive microwave remotely-sensed data is a function of surface soil moisture, surface roughness, vegetation cover and changes in dielectric constant related to soil texture. It has been used successfully in a number of applications but the use of passive microwave measurements for soil moisture mapping is limited by the coarse resolution. As of the date of this publication, the following passive microwave sensors have been identified as having potential for estimating soil moisture: the Advanced Microwave Scanning Radiometer (AMSR-E) deployed on the NASA Aqua platform in 2003, the Soil Moisture and Ocean Salinity (SMOS) mission planned for launch by the European Space Agency (ESA) in 2007, and the NASA hydrospheric states (HYDROS) mission planned for launch in 2009 (Moran *et al.*, 2004).

Considered to be the most promising approach to estimating soil moisture at a fine resolution, active microwave sensing applications for measuring surface soil moisture are becoming increasingly common. Synthetic aperture radar (SAR) transmits a series of pulses as the radar antenna traverses the scene, but a significant limitation of this technique is that the sun-synchronous satellites provide only weekly or even longer repeat coverage for the same orbital path (Moran *et al.*, 2004). In addition, the current satellite-based SAR sensors are configured with only a single wavelength (C- or L-bands) and, in some cases, with a single incidence angle. Furthermore, RADARSAT and ERS (European Remote-Sensing Satellite) SAR sensors offer only single polarization (either horizontal (HH) or vertical (VV)).

Unfortunately, methods of estimating soil moisture using active or passive microwave data are still not widely applied for a number of reasons. The availability of ground-truthing data at an appropriate scale is limited, soil moisture content is vertically heterogeneous (Carlson *et al.*, 1995), and account needs to be taken of the confounding effects of vegetation cover (Jackson, 1997) (especially problematic when only a single-wavelength dataset is available) and surface roughness (Jackson *et al.*, 1997). The latter factor is particularly important because the sensitivity of radar backscatter to roughness can be much greater than the sensitivity to soil moisture.

Soil moisture cannot be estimated from single-wavelength, single-incidence angle, single-pass SAR data without information about the surface roughness (Moran *et al.*, 2004) and topography. A potential operational application using single-wavelength, multi-pass SAR images, is detection of change instead of absolute values of soil moisture content. However, the assumptions underpinning this application do not hold for cultivated crops where roughness and vegetation density can change dramatically between passes. Furthermore, images must also be acquired with the same sensor configuration to avoid the need for topographic corrections due to variations in incidence angle and image orientation (Moran *et al.*, 2004).

In their review of estimating soil moisture with satellite-based radar, Moran *et al.* (2004) propose that research and development should focus on: mapping roughness or determining the soil moisture in densely vegetated sites; further development of theoretical backscattering models; further development of data fusion techniques to determine subpixel variability of passive-derived soil moisture with the finer resolution active microwave data; and decreasing SAR imagery prices. They go on to state: "*Once all the pieces are in place, a primary obstacle to operational application will be sheer technological complexity involved in the implementation of a remote sensing/ backscatter model/SVAT system by hydrologists, geospatial analysts, and watershed managers. This could be accomplished by a commercial venture or government space agency dedicated to providing information of the resolution, coverage, and frequency required … though no such operation currently exists. If decentralized, a … [user] would need to employ computer-savvy personnel who could make the high-technology, high-science components work smoothly with minimal technical support.*" (Moran *et al.*, 2004).

4.4.4 Identifying flow pathways, runoff components and water age using tracers

Different tracer techniques have allowed deeper insights to be gained into the origin, flow pathways and residence times of water in basins of different sizes. Chemical tracers (e.g. dissolved silica, different anions and cations) enable surface and subsurface runoff components to be differentiated during storm events (e.g. Pinder & Jones 1969; Hooper 2001) and the use of environmental isotopes make it possible to estimate the age of the water. Thus, on the event time scale it is possible to quantify the contribution of event water and pre-event water (water that was stored in the basin before the event) using oxygen-18 (^{18}O) and deuterium (^{2}H) (e.g. Sklash & Farvolden, 1979). At a longer time scale it is possible to use a variety of environmental tracers and pollution tracers (e.g. CFCs, SF_6) (Kendall & McDonnell, 1998) to determine the residence time distribution (often characterized by the mean residence time) of water in different hydrological storages, for instance in groundwater bodies and lakes (Maloszweski & Zuber, 1982). Therefore, the mixing of different water compartments can also be defined using chemical and isotopic tracers, for example enabling crucial surface water/groundwater interactions and groundwater recharge to be investigated.

Artificial tracers, such as different dyes, salts, bacteria, viruses, small particles, micro-chips and DNA parts (Kaess, 1998), are very useful for investigating flow and transport processes at the surface and in the subsurface (e.g. Mosley, 1982; Hornberger *et al.*, 1991). An example of the use of an artificial tracer, naphthionate, to identify sources of runoff is shown in Fig. 4.2. The advantages of artificial tracers are that they provide absolute proof of a flow path and allow the direct measurement of flow and transport parameters, but the technique is limited to application at relatively small spatial and temporal scales.

Tracer methods have been applied at different scales: all kind of tracers have been used for plot scale investigations (McDonnell, 1990; Hornberger *et al.*, 1991), whilst investigations at the basin scale (micro- and meso-scale) generally use natural tracers (Pearce *et al.*, 1986; Uhlenbrook *et al.*, 2002). Recently, environmental isotopes have

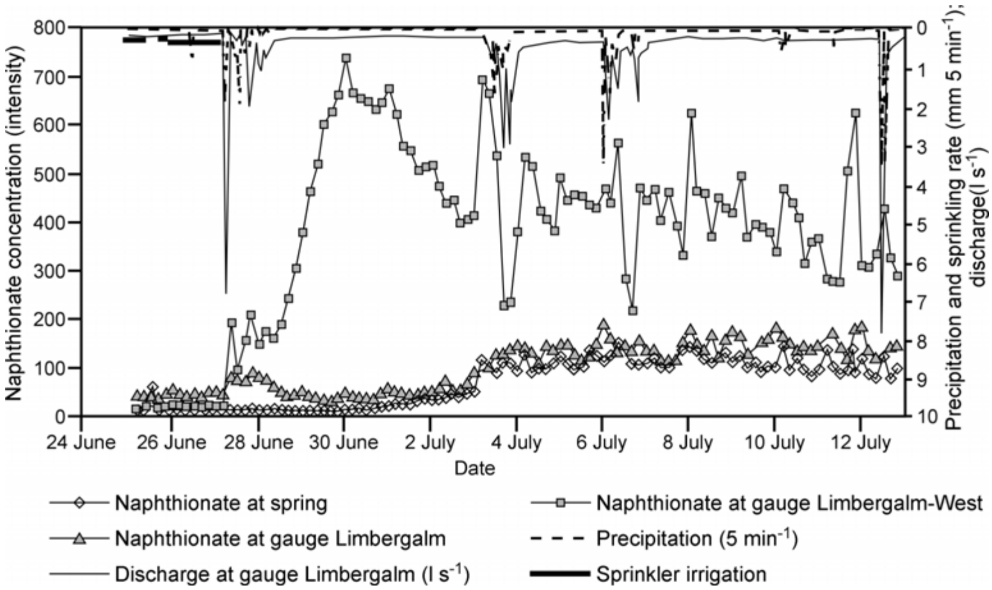

Fig. 4.2 *Results from sprinkler application of an artificial tracer (naphthionate) to a site at the Loehnersbach catchment, Austria, showed that water in the saturated area and springs originates from the upper steep hillslope area (after Tilch et al., 2006).*

also proved useful at the macro-scale; Bosilovich & Schubert (2002) used environmental isotope data in GCM simulations to trace the route of evaporated moisture from the land surface through the atmosphere and then as precipitation at different locations.

4.4.5 Surface discharge

The prediction of surface discharge is critical for a whole range of applications. Basin surface discharge represents a basin-integrated hydrological flux in that the response of the basin area should drain through one point (or outlet). In practice hydrological basins are typically defined according to surface topography, which may or may not correspond to areas where subsurface controls may dominate fluxes. Nevertheless, surface discharge provides water for abstraction, maintains many surface ecosystems, and periodically brings devastating floods. Basin discharge is usually estimated through the installation of a gauging device such as a weir or flume. These relate the water level upstream of the device to the discharge through it. As the derivation of the basic form of this relationship is based on idealized fluid mechanics, the coefficients of the relationship for each device must be calibrated either in the laboratory or *in situ*. *In situ* methods are point-based sampling of the water surface that rely on well defined channel boundaries to confine the flow. Difficulties arise in attempting to gauge surface discharge in wetlands, braided rivers and flood plains, which are spatially complex with both vast diffusive flows and narrow confined hydraulics (Alsdorf & Rodriguez, 2005).

Collection of surface water discharge data currently suffers from many problems including inadequate data quality, particularly in developing countries, reductions in hydrometric networks almost universally in all countries, and poor access or restrictions on collected data. However, remote sensing measurement strategies are being developed that will provide surface water data at relatively low cost and enable data to be collected and disseminated without regard to political boundaries. Remotely-sensed observations of flooded areas and water levels have the potential to directly measure the runoff component of the global hydrological cycle (Brakenridge *et al.*, 2005).

<u>Remote sensing of surface discharge</u> A variety of surface water parameters can be estimated using remote sensing. Surface water extent can be measured using either optical or microwave wavelengths, although the usual problems are encountered, such as: cloud contamination; either poor temporal or spatial resolution; and, for optical sensors, the inability to penetrate vegetation canopies that may be flooded. Shorter wavelength SAR-based estimates of surface water extent are confounded by difficulties associated with wind-roughening of the water surface, but this may be resolved by multiple passes if there is little change in the surface conditions (Alsdorf & Rodriguez, 2005). At longer L-band wavelengths, radar energy is unaffected by wind or wave action on the water surface and is capable of penetrating vegetation canopies, allowing imaging of the underlying flood water surface. While there has been some success using passive microwave sensors to determine the area of inundation, their low spatial resolution means that they do not work well in environments where small changes in water heights yield little change in surface area but significant changes in flow (Alsdorf & Rodriguez, 2005).

Both profiling and imaging methods using radar, SAR or optical wavelengths are available for measuring surface water elevations. Radar altimetry of the kind available with TOPEX or EnviSat can produce height resolutions for large lakes and waterbodies of typically 50 cm. Inland water measurements from radar altimetry have a much degraded resolution because of reduced pulse averaging and differing echo shapes, but spaceborne LIDAR, such as the GLAS instrument on ICESAT, can provide accuracies of 3 cm for smaller water bodies (Alsdorf & Rodriquez, 2005). Satellite-based river measurements use *gauging reaches*, rather than gauging stations, to measure changing water surface areas within carefully defined river reaches. The most suitable reaches are those where discharge changes are accompanied by flow width and water surface area expansions and contractions. The sensitivity of this method depends on reach channel and flood plain morphology, on sensor spatial resolution, and on the precision and accuracy of the classification of water area (Brakenridge *et al.*, 2005). Space-based surface water observations are not as precise as gauging station data, and the relationship of reach surface area to discharge may be affected by errors such as hysteresis (Brakenridge *et al.*, 2005).

The problems with existing satellite-based technologies for measuring surface water are summarized by Alsdorf & Rodriguez (2005) as:

– Inability for optical wavelengths to penetrate clouds, smoke or vegetation. This is a particular problem in the tropics due to the prevailing vegetation and atmospheric conditions.
– Poor spatial resolutions.
– Conventional radar and lidar altimetry misses water bodies between orbital tracks.
– Current operating radar altimeters are not built to sample small bodies.
– SRTM (Shuttle Radar Topography Mission) errors over water surfaces are large and temporal resolutions are poor.
– Interferometric SAR will not work over open water and requires special hydrogeomorphologies of flooded vegetation.

4.5 QUANTIFICATION OF PARAMETERS AFFECTING GROUNDWATER

Conventional subsurface sampling techniques for characterizing subsurface properties or for monitoring hydrological processes typically involve collecting “point” samples, or drilling a borehole, wherein soil samples are retrieved for further analysis, or borehole logs are collected. These methods can be costly, time-consuming, and invasive, potentially disturbing the *in situ* conditions of interest and exposing humans to hazardous contaminants. Because these measurements are typically sparse and often associated with a very localized measurement support scale, they often do not provide sufficient information about field-scale hydrogeologic heterogeneity or processes that is needed to guide water resource management or environmental remediation. Geostatistical methods are often used to interpolate between point-based methods (see for instance Issaks & Srivastava, 1989). Many powerful stochastic methods have also been developed to quantify uncertainty associated with estimating parameters or predicting processes in the face of sparse datasets and modelling uncertainties (refer to Chapter 5); a good review of groundwater stochastic groundwater hydrology is given by Rubin (2003).

By combining measurements obtained using geophysical and hydrological methods, accurate subsurface characterization and monitoring can potentially be achieved with high temporal and spatial resolution and over a range of spatial scales. The existence of such potential is intuitively appealing: geophysics offers a view of the subsurface in a minimally invasive manner and from a perspective other than hydrogeology, and thus, when reconciled with the hydrogeological perspective, can help obtain a more coherent image.

As described by Hubbard & Rubin (2005), successful integration of geophysical and hydrogeological datasets to provide quantitative information about hydrological properties and processes represents a recent breakthrough in hydrogeological site characterization. Just as medical imaging technology provides dense information and has reduced the need for invasive surgery, the value of hydrogeophysical characterization lies in the extensive spatial coverage offered by geophysical techniques and in their ability to sample the subsurface in a minimally invasive manner. Time-lapse imaging, especially using high-resolution cross borehole data, has illustrated the potential of hydrogeophysical methods for elucidating dynamic subsurface processes. An example of the utility of combined hydrogeophysical and tracer approaches is shown in Fig. 4.3.

4.6 THE CRYOSPHERE: SNOW AND ICE

The cryosphere (snow and ice) is an important, and often neglected, component of the hydrological cycle. It is one of the most responsive components of the hydrological cycle to climate change, yet it is also an important source of runoff to many arid and semi-arid regions of the world. This section reviews methods for monitoring and modelling the snowpack, ranging from simple meteorological-based models to those that attempt to simulate processes within the snowpack, and identifies future research needs.

4.6.1 Monitoring snowpack mass balance

In mountainous regions the snowpack plays a major role in the timing of flows by storing water in winter and delivering it during warm periods. It is therefore necessary to monitor the snowpack in order to estimate the snow water equivalent (SWE) available during the melting period and also to determine the snowpack spatial distribution in relation to altitude, which controls how liquid water will be delivered as a function of meteorological conditions.

Measurement of snow depth is the oldest and simplest way of monitoring the snowpack. It is generally done using graduated rods, which are periodically read during winter (every day, week or month). The method is quite cheap, easy to apply and multiple rods can be used to estimate the spatial variability of the snowpack. The spatial representativity of a single measurement is a subject of discussion (e.g. Sommerfeld *et al.*, 1991), but authors generally agree that is quite small (1–100 m), although this depends on the topography and aspect. This method has some additional disadvantages. Firstly, it does not allow the processes affecting snow accumulation (precipitation *vs* snowdrift) and decrease (melting *vs* settlement) to be identified. Thus, the rods need to be carefully located by choosing places representative of a large area and which are not affected by snow transport. Secondly, the method is not

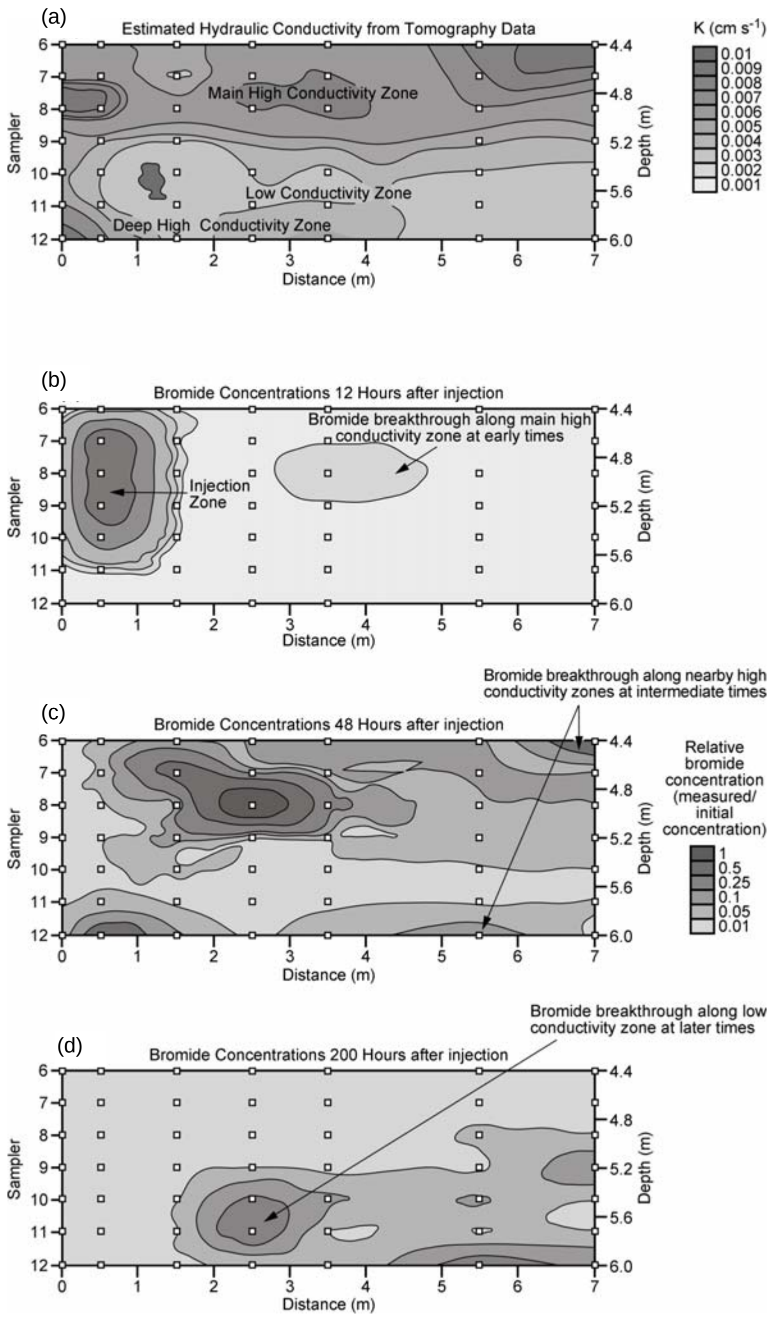

Fig. 4.3 *Comparison of: (a) the log hydraulic conductivity (K) values estimated using tomographic data with bromide relative concentration, at (b) 12 h, (c) 48 h and (d) 200 h after injection in a sandy Pleistocene aquifer, Virginia, USA, suggests the utility of tomographic methods for providing information necessary for flow and transport modelling. Depth is depth below surface in metres, referenced to mean sea level. Visual examination suggests the plume travels quickly along the high-hydraulic conductivity zone(s) interpreted using the tomographic data and, after long times, resides in the low-conductivity zone (after Hubbard et al., 2001).*

applicable for monitoring in large basins or high mountainous basins (which are difficult to access in winter). In these situations an alternative approach is automatic snow depth measurement (using ultrasonic sensors), but the cost of each monitoring station does not allow numerous measurement points. Lastly, for hydrological purposes snow depth measurement provides an estimate only of snow accumulation, but not directly of the SWE, the most useful information for water resources applications. To estimate the SWE the snow density, which ranges from 50 kg m^{-3} for fresh snow to 600 kg m^{-3} for very old seasonal snow, needs to be known. Techniques are presently under development to directly estimate point values of SWE, for instance by measuring the attenuation of cosmic or radioactive radiation by the snowpack.

In the last 20 years alternative methods for snowpack monitoring have been developed, many based on remote sensing. Snow cover may be monitored using satellite images in visible and infra-red channels (e.g. NOAA/AVHRR) where the snow signature is very distinct. Depending on the satellite the snow cover can be monitored at a high spatial (e.g. Landsat, Spot satellites) or temporal (National Environmental Satellite, Data and Information Service (NESDIS) satellite) resolution, covering areas as large as the northern hemisphere. In smaller catchments snow cover can be estimated directly from photographs taken from aeroplanes or from the ground and projected onto a numerical terrain model and ground radar can also be used to detect snow cover. Remote sensing can only measure snow cover and not depth, but passive microwave measurements allow the SWE to be estimated with Special Sensor Microwave Imager (SSM/I) observations (e.g. Goodison & Walker, 1995; Armstrong & Brodzik, 2002). However, this method is only applicable for large homogeneous snow fields and requires ground measurements for calibration to take account of liquid water in the snowpack.

To realize improvements in large-scale snowpack monitoring there is a need to combine measurements from different sources, combining ground information and satellite data. Numerous methods developed for glacier monitoring, such as seismic reflection and portable radar, synthetic aperture radar (SAR) imagery and laser and radar altimetry, could be adapted to snow monitoring. This would enable remote sensing of the snowpack from air and space and provide information that is currently lacking on high altitude and desert regions.

4.6.2 Snow accumulation and distribution

One of the main difficulties in characterising the snowpack is to define the relevant spatial scale. Variations in the snowpack occur at a range of scales from metres to kilometres, depending on the vegetation type, the local wind patterns and the micro- and meso-topography (Marsh, 1999). The two main causes of variability in snow accumulation are snow transport by wind and interception by trees. In the last 20 years considerable effort has been expended to develop physically-based models to increase knowledge of these processes.

Snow transport by wind significantly affects SWE as a result of snow redistribution. Wind also increases the snow sublimation rate (normally 15–40% of snowfall depending on climate, wind fetch and vegetation) by increasing ventilation of the snow crystals (Pomeroy *et al.*, 1998) and modifying the snow chemistry. Recently models have been developed which take into account the ability of snow to be transported and calculate snow distribution as a function of wind fields. These models particularly focus on the effect of vegetation type on snow redistribution (e.g. Essery *et al.*, 1999a) and generally show that the regional SWE can be significantly lower than the snowfall amount during the accumulation period because of greater sublimation rates (see Pomeroy & Gray (1994) and Liston & Sturm (1998) for reviews of these studies). Even if these models work reasonably well at large scales, they need to be improved in order to simulate snow distribution in complex mountainous terrain.

Snow interception by the forest canopy also has a major effect on the snowpack, particularly for non-deciduous forests. The canopy interception efficiency depends on vegetation characteristics (leaf shape and type, branch structure, tree size), snow type (wet, dry, blown snow) and meteorological conditions (which affect how long snow remains in the canopy). In high-latitude boreal forests snow can remain on trees for weeks and a large part of the intercepted snow sublimates to the atmosphere. Many studies have been conducted on single branches or trees to understand the factors affecting snow interception and retention by the canopy. Physically-based models have been developed to extrapolate the results from these small-scale investigations to forested areas (see Pomeroy & Gray, 1994 and Hedstrom & Pomeroy, 1998; for reviews). Although there has been considerable effort in developing these models, they are currently limited to particular forest environments (mainly boreal forest) and need to be validated for other tree species and mixed forests.

4.6.3 Water routing in the snowpack

Water routing into and out of the snowpack is a subject of particular interest to hydrologists. Water percolation in snow is especially complex for a number of reasons: water can melt and refreeze in the snowpack during percolation; percolation is spatially heterogeneous and follows preferential and non-uniform flow paths; ice layers strongly modify downward water flows; and snow grains quickly evolve when they are in contact with liquid water. Numerous studies have been conducted to improve understanding of these complex processes, particularly to describe the formation of preferential flow paths and the formation of ice layers, but they are often qualitative due to limitations in observation capacity. Some sophisticated observation tools (such as radar or ultrasonic) have been proposed but have not yet been developed for this application. Some models, such as finite-element models, have also been used to describe water percolation in snow allowing a better understanding of ice formation in the snowpack and its effect on water flow. However, the models assume that the snow is isothermal and do not take account of the spatial distribution of flow paths (which requires a complete three-dimensional description of the snowpack) (Marsh, 1999).

4.6.4 Modelling snowpack energy and mass balance

For more than 30 years models have been used to estimate snowpack evolution from measured meteorological parameters. In temperate and high latitudes the snowpack varies seasonally: increasing in the cold season (accumulation) and disappearing in the warm season (melting). For several months (the exact duration depends on latitude, elevation, aspect and climatic conditions) the soil is isolated from the atmosphere by the overlying snow and moves towards thermal equilibrium with the base of the snowpack. As soil is generally warmer than snow, a weak and fairly constant flux (0 to a few W m^{-2}) transfers energy from soil to the snowpack. This flux is generally insufficient to cause snowmelt and its main effect is to maintain a mean vertical snow temperature gradient oriented toward the soil.

The main energy exchanges governing the snow energy budget are between the atmosphere and the snowpack. Consequently the simplest models of snowpack evolution (the degree-day models) compute the daily evolution of the snowpack from air temperature and precipitation data. It has been widely demonstrated that this method provides good results for small catchments with a limited elevation range. However, the degree-day factor requires calibration for individual sites and typical values reported in the literature range from 1 to 5 mm °C day^{-1}. The dependence of snowmelt on air temperature in degree-day models is a large simplification of the surface energy budget, which also includes downward and upward visible and infrared radiation, turbulent exchanges and precipitation energy transfer.

Hence, in the 1980s, more sophisticated models were developed that generally include several snow layers and simulate internal processes within the snowpack (such as, percolation and water storage by snow). The surface energy budget (which governs snowmelt) is calculated in the models from meteorological measurements that include visible and thermal radiation, air temperature and humidity, wind speed and precipitation. The main difficulties are estimating the snow albedo (which is dependent on the snow micro-structure) and the snow surface temperature (which has complex interactions with the components of the energy budget). The model time step needs to be small enough (15 min to 3 h) to model processes such as water refreezing in snow. These models have a number of advantages over the degree-day models. They are more accurate and require very limited calibration when transposed from one basin to another. Particular climatic conditions and their effect on the snowpack (for instance a high sublimation rate) can be taken into account and spatial variability of the snowpack in a basin (due to aspect or elevation, for example) can also be modelled. However, they are very data intensive and complex to operate so that in many instances they cannot be applied because insufficient data are available.

Consequently, models of intermediate complexity have been developed more recently. These are based on the degree-day approach but take into account additional information such as wind speed (which affects the degree-day factor), the spatial pattern of radiation (which results in increases or decreases in melting rate depending on slope aspect) or measurements of snow cover area (derived from remote sensing observations). Several authors have reviewed these models and compared the performances of models with different complexities (e.g. Essery *et al.*, 1999b; Schlosser *et al.*, 2000). The Snow Model Intercomparison Project (Etchevers *et al.*, 2003) systematically compared 26 snow models with a wide range of complexities depending on their application (e.g. climate, hydrology, avalanche hazard forecasting). Comparisons of model simulations of SWE and net long wave radiation with local observations for three different sites are shown in Fig. 4.4. The complexity of the models was classified on a scale from 1 (very simple model with a uniform snow layer and a simplified energy budget) to 4 (sophisticated multi-layer model that simulates internal snow processes). An interesting result is that even simple models (with a complexity of 2) can give very good results for the different sites (low RMS when model outputs compared with observed data). However, some models provide a good simulation of the snowpack mass balance but a poor simulation of the net long wave radiation. Only the most sophisticated models are able to simulate both the snowpack mass balance and surface energy fluxes with good accuracy. Hence the choice of snow model depends on the nature of the application and the degree of accuracy that is required.

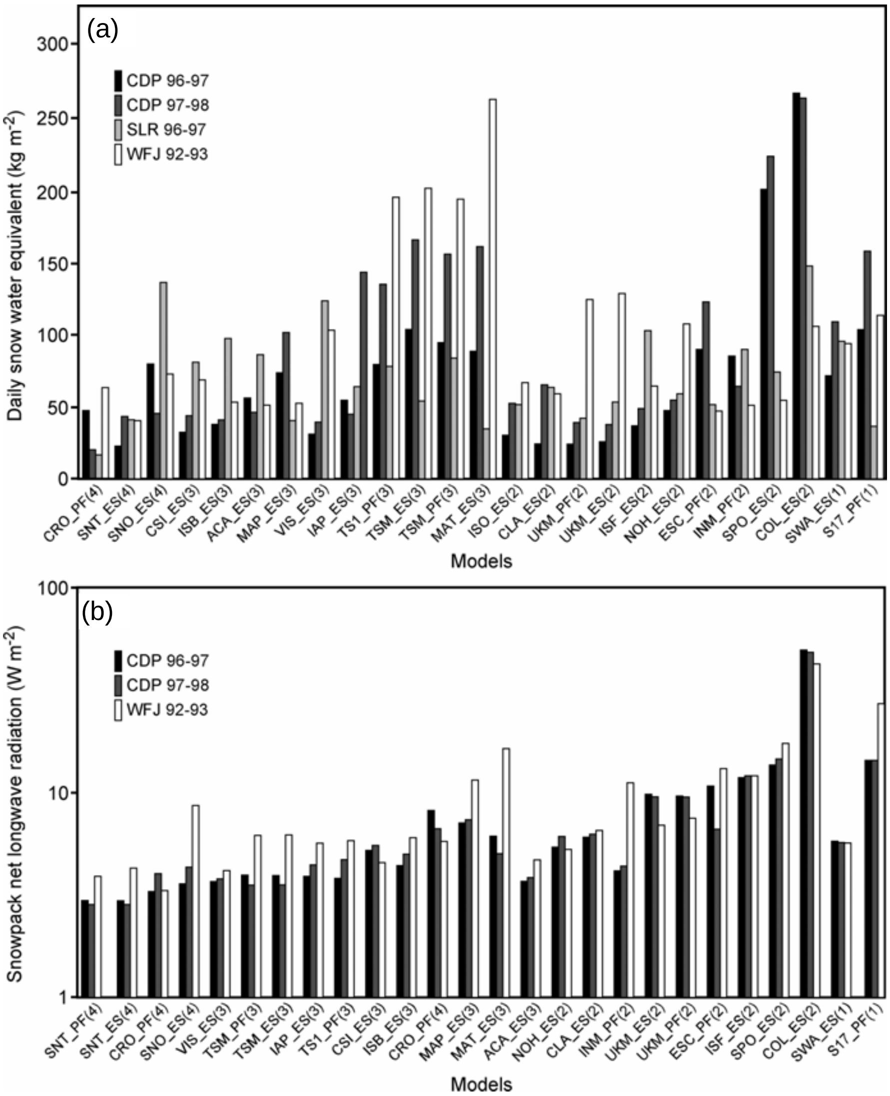

Fig. 4.4 *Results from the Snow Model Intercomparison Project (SnowMIP): root mean square (RMS) error for (a) daily SWE and (b) snowpack net long wave radiation for three sites: Col de Porte (CDP), France; Weissfluhjoch (WFJ), Switzerland; Sleepers River (SLR), Vermont, USA, for a winter season. The RMS error has been calculated using local observations of SWE and snow surface temperature (the reference) and the simulated SWE and snowpack net long wave radiation from 26 different snow models (acronyms on x-axis). The model complexity ranges from 1 (very simple) to 4 (the most sophisticated models that currently exist) and is indicated in brackets after the model acronym. (Figures adapted from Etchevers et al., 2003).*

4.6.5 Snow surface energy fluxes

A further aspect that needs to be considered in modelling snowpack energy is the surface energy fluxes which are dominated by radiative fluxes. The main energy source is short wave solar radiation which influences snow cover as a function of altitude, aspect and latitude. Net short wave radiation depends on the snow surface characteristics (grain type, liquid water content, dust content) and is higher for large grains and old snow. Long wave radiation emitted by snow is controlled by the snow surface temperature. Turbulent fluxes influence snow surface temperature and energy balance but are poorly characterized because they are difficult to measure. The boundary layer above snow is generally very stable because the snow surface temperature is generally much lower than in overlying air. Moreover, no complete theory exists to describe turbulent exchange in mountainous terrain and/or above snow fields or glaciers.

Methods based on eddy correlation or aircraft measurements have been tested to measure turbulent fluxes and their spatial variations at the snow surface. Substantial work remains to validate these methods for snow surfaces, which are generally complex (due to topography) and require the spatial variation of incoming radiation (due to surrounding slopes or presence of a forest canopy) to be taken into account. Particular efforts have focused on the role of local scale advection over snow patches. Authors have shown that snow patch pattern has an important effect on the spatially-averaged snow melt rate, and can result in increases of snow melt of up to 30% (Liston, 1995). Future developments will require these studies to be transposed to mountainous terrain and consideration of the relative importance for the surface energy budget of energy released or consumed in biological and chemical processes (Jones *et al.*, 1993).

4.6.6 Snow–vegetation interactions

Interactions between snow and vegetation are particularly important in forested areas. Snow cover can modify the lifecycle of forests, whilst trees have a large influence on snowpack development. Snow intercepted by trees may sublimate or fall to the ground after some days modifying the distribution of snow, from a very thin cover close to tree trunks to a deeper snowpack in open areas between trees. Energy exchanges at the snowpack surface are strongly reduced (Hardy *et al.*, 1997) for a number of reasons (intercepted incoming radiation, reduced wind speed, infra-red radiation emitted by trees, changes in air humidity and temperature in forests). The resulting global budget is also changed: in general the average SWE is lower and snow melt is slower in forested areas. In recent years there has been considerable effort to monitor the snowpack under trees (e.g. Hardy *et al.*, 1997; Sellers *et al.*, 1997). Using these field measurements, physically-based models (particularly adapted radiative models that take account of multiple reflections between the snow and the canopy bottom) have been developed to estimate changes in the snowpack induced by forest and the feedback between the snowpack and the trees. Although these studies explain the evolution of the snowpack beneath a forest reasonably well, there is a need to develop such models so that they can be applied to other types of forests.

4.6.7 Future outlook for hydrology of the cryosphere

Studies of the cryosphere have made significant progress in the last 20 years. New methods have been developed to measure physical processes specific to the snowpack and ice sheets for a large range of spatial scales (from grain size to hemisphere scales).

At the same time comprehensive models have been developed in order to better understand complex phenomena like snowpack interactions with the vegetation canopy, snowdrift and turbulent energy exchanges with the atmosphere. In the future, hydrologists and other scientists (e.g. meteorologists, climate specialists) will need to address the major challenges of monitoring and modelling the cryosphere for at least two reasons. Firstly, the snowpack is a major component of the regional water resources in several parts of the world. It stores water in the upper reaches of basins and strongly controls the discharge of mountain rivers. In mountainous regions, such as the Himalayas or the Andes, the cryosphere may be the main water resource for very large populations during the dry season. Secondly, the climate and cryosphere are closely-coupled. Consequently the cryosphere is one of the first indicators of climate change and plays an important role in the climate system because of the feedbacks between snow cover and sea-ice albedo and climate. Future advances are expected in remote sensing data use and development of new satellite sensors specific to the cryosphere. For local hydrological applications, specific field measurements will continue to increase understanding of complex physical processes and to improve current models and develop new modelling approaches.

4.7 MEASUREMENT FOR WATER QUALITY

The measurement of water quality includes issues of field measurement and appropriate sampling strategies and of sample analysis. Water quality ranges from physical components (for instance, pH, electrical conductivity, redox-potential, temperature) to numerous chemical components (including cations, anions, hydrocarbons, and naturally occurring and synthetic organic compounds). A further level of complexity in water quality assessment is the occurrence of different forms of chemical components, for example nitrogen may be present in water as inorganic nitrate (NO_3^-), nitrite (NO_2^-) and ammoniacal-nitrogen (NH_4^+), organic nitrogen and in particulate and dissolved forms. It is often important to distinguish between different chemical forms because they have differing toxicological and ecological effects. Methods of water analysis are standardized and well-documented (see for example, Clesceri *et al.*, 1999; methods published by the Water Quality Technical Committee of the International Organization for Standardization (ISO); UK Standing Committee of Analysts Methods for the Examination of Waters and Related Materials published by HMSO and the Environment Agency) and the quality of analysis can be assured through quality control procedures, laboratory accreditation schemes and the use of certified reference materials. The key questions for the hydrologist in relation to water quality analysis are normally which parameters to measure and the costs of analysis.

With regards to direct field measurement, at the moment a few water quality variables can be readily measured *in situ* with probes (e.g. pH, conductivity, turbidity, nitrate) and costly continuous analysers, but the determination of most water quality variables requires the laboratory analysis of water samples. Time delays between sample collection and analysis can be problematic as the composition of water samples may alter during storage and transport, particularly for nutrients and alkalinity. The costs and time involved in laboratory analysis of water samples often necessitate the design of a water sampling strategy to derive the maximum information from the minimum number of samples collected in order to meet the study aims.

Protocols for the collection of representative samples from surface and groundwaters are outlined in UNEP (1992) and sampling methods published by the Water Quality Technical Committee of the International Organization for Standardization (ISO). Sampling water from the vadose zone is more complex as the physical action of sample collection may perturb soil water flow pathways, in turn affecting the composition of soil water. One of the simplest methods for soil water sampling involves collecting water from inclined troughs inserted into an exposed soil face. However there are uncertainties regarding how representative these samples are of the actual soil water composition because of the disturbance to soil water flow pathways by the introduction of the troughs. A widely used method of soil water sampling involves porous ceramic cup samplers or tension lysimeters. The principle behind these samplers is the application of a negative pressure to a sealed container, attached to a porous cup (normally composed of a ceramic material but sometimes made of another material, such as teflon) buried in the soil. Water in the soil surrounding the cup flows into the container for as long as the negative pressure in the cup exceeds the external soil moisture potential. The limitations of this method for collection of representative samples of soil water are well-documented (e.g. Barbee & Brown, 1986; Debyle *et al.*, 1988; Morrison & Lowery, 1990; Grossmann & Udluft, 1991; Wu *et al.*, 1995; Hatch *et al.*, 1997) and include: variations in soil volume sampled by the cup; macropore flow bypassing the cup; leaching of contaminants from the cup material into the water sample; adsorption of cations by the cup material; the effect of the initial vacuum applied to the cup on the sample collected; alteration of sample composition during storage within the cup; blocking of the cup pores and failure to collect water in dry soil conditions. Other means of soil water sampling include zero tension lysimeters (collection of drainage from an isolated block of soil, but the lysimeters are costly to construct) and the destructive method of centrifuging soil samples to obtain soil water.

As well as the sampling and analysis of water samples for water quality, hydrologists are often concerned with the analysis of the resulting water quality data to estimate fluxes of solutes and particulates required for basin biogeochemical cycling or diffuse pollution studies. The calculation of fluxes in water is complicated and subject to uncertainty as water quality measurements are normally only available for discrete samples in time and space, and relationships often exist between the concentration of a water quality variable and flow through the processes of flushing, dilution and hysteresis, for example as the store of a chemical is depleted over the course of a storm event. Consequently hydrologists continue to devise water quality sampling programmes and methods of data analysis to reduce such uncertainties and improve the accuracy of estimates of solute and particulate fluxes (e.g. Reynolds *et al.*, 1990; Robertson & Roerisch, 1999).

In summary, water quality measurement is currently in a state of transition from discrete point measurements of individual samples in laboratories to continuous real-time, spatially distributed *in situ* measurements coupled with flows. The latter are only beginning to be realized in a limited number of basins, predominantly in developed countries (e.g. Evans *et al.*, 1997). There is also now a greater emphasis on integrated and holistic measures of water quality. Measurements of the physical and chemical properties of water only provide an instantaneous picture of water quality which can change dramatically over short periods of time in dynamic surface water systems.

Consequently more biological measures are being developed (see Section 6.4) to provide temporally integrated measures of water quality. Nevertheless, instantaneous physical and chemical measures of water quality are still of particular relevance to hydrologists as tools for furthering understanding of flow pathways and the processes by which contaminants enter runoff.

4.8 MEASUREMENT FOR LAND EROSION

Land degradation is a complex phenomenon which involves physical, chemical and biological processes in combination with climatic (e.g. rain, wind, temperature), natural (e.g. relief, vegetation, soils) and anthropogenic factors (e.g. land use, economy, political strategy). Water erosion represents one of the gravest and most widespread forms of land degradation. Measurement techniques for land degradation are numerous, various and, in most cases, not standardized. The choice of technique depends on the process studied, the observation scale and the study objective. However, techniques can be grouped in two categories (Lal, 2001): the measurement of current soil erosion and the measurement of accumulated historical erosion.

In the first category, different techniques are used (Lal, 1994) according to the type of erosion studied (e.g. water, wind, river, splash effect, sheet, gully, bedload, landslide) and the study scale (plot, field, catchment, basin). Some of the most widely used traditional techniques are runoff and erosion plots and surface survey. The plots used in numerous studies are generally of different size and design, making it problematic to compare results because the data obtained are often technique-dependent (Lal, 2001). In addition, since plot studies are used at small and more or less homogeneous scales, it is difficult to extrapolate the results to larger scales with greater spatial heterogeneities. Today increasing use is made of “intelligent” techniques for measuring soil erosion and this tendency is likely to accelerate in the future with the development of nanotechnology. Examples of new techniques are the optical spectro-pluviometer for the study of raindrop erosivity, nuclear gauges for suspended sediment measurement, vortex flume and electromagnetic devices for bedload transport, and laser- and optical-based sensors for suspended sediment monitoring.

The second category of techniques concerning the measurement of accumulated historical erosion, focuses on the acquisition of data over longer periods (decades). One of the most widespread techniques is the use of fallout radionuclides to estimate rates of erosion and deposition in agricultural basins (Zapata, 2003). Caesium-137 (^{137}Cs) has been widely used to identify the sources and extent of soil loss (Walling, 2003) because of its good marking of fine particles, its long half-life duration (about 30 years) and the relative ease of measurement. Other environmental radionuclides, such as lead-210 (^{210}Pb) and beryllium-7 (^{7}Be) are also used in the study of soil erosion, either individually, or in combination with ^{137}Cs. Whilst the use of radionuclides can provide data on the spatial distribution of erosion and deposition over long time periods, application of the technique is currently expensive and requires an interdisciplinary team and well equipped specialized laboratories.

Future developments in the measurement of land erosion are expected to revolve around the application of new technologies, such as remote sensing and GIS, that can provide and manage the fine spatially-distributed information (e.g. variability in land uses/cover) required to simulate and predict erosion and land degradation.

4.9 EMERGING TECHNIQUES FOR HYDROLOGICAL MONITORING

4.9.1 Microsensors and networked infrastructure

As indicated earlier, there are several emerging techniques and sensor types for hydrological characterization, including hydrogeophysical and tracer approaches, and microsensor and remote sensing methods. Some of the traditional measurement tools used in hydrology were designed for purposes other than measuring hydrological variables and parameters. An excellent example is the field of remote sensing where numerous satellite-based sensors are used to provide regional scale hydrological parameters for which they were never intended. Advances in technology will naturally facilitate advances in hydrological measurement, but it will be the development of instrumentation specifically designed for hydrological monitoring, which is scale specific and cost effective, that will lead to the greatest progress in the field of hydrology.

The development of networked infrastructure as a platform for collecting measurements is an emerging trend in environmental science. Distributed systems composed of smart, small and wireless sensors that are embedded in the natural environment offer the potential to match the spatial and temporal density of the measurements with the scale of the problem. The ability to deploy, at a reasonable cost, hundreds or thousands of sensors has the potential to revolutionize system characterization and monitoring.

Efficient use of microsensors involves both evolutionary and revolutionary advances in miniaturization, integration, communication technology, and energy management. MEMS, or micro electro mechanical systems, are devices that are embedded into semiconductor chips and which can be used to detect fluid pressure, pH, temperature, and vibrations. The development of particular chemical sensors is underway. Smart dust (also called motes) is an extension of the technology that includes on-board power and radio communications. Operation systems enable the motes to “talk” with each other, and to form a measurement network. Each group of sensors ultimately communicates with a laptop base station, where the data are stored in a database. The devices are essentially inexpensive, self-contained, battery-powered computers with radio links that enable them to wirelessly self-organize into a network and to communicate with one another (at distances of ~300 m). Motes range in size from a micrometre to a millimetre, although current research is underway to reduce mote size to that of a sand grain or dust speck. Because of their miniscule size and wireless capabilities, they can be installed almost anywhere with minimum effort.

Microsensor technologies are currently being explored for use within the atmosphere (e.g. Manobianco *et al.*, 2003) and at the Earth’s surface (Polastre, 2003) to assist with weather forecasting and ecological investigations, respectively. Microsensor network performance has increased by approximately two-fold every 18 months, and enhancement efforts continue on several fronts. Much of the current microsensor network research focuses on development of particular sensors, on ultra low-power systems, on methods to handle the plethora of data that is often sent back to traditional servers, and on development of adaptive feedback systems for particular applications. The accuracy of the sensors can be improved by deploying many inexpensive sensors redundantly so that the signal to noise ratio of the ensemble data and the reliability are improved. Probe lifetime may vary significantly depending on factors such as power storage/consumption, sampling rate, communication frequency, and on-board data storage and processing. These characteristics will influence the quantity of data available for assimilation into investigations of hydrological systems, as well as the predictions generated using such data.

4.9.2 Tracers and hydrogeophysical approaches

In the last decade, many advances associated with near-surface geophysics have been made that facilitate the use of geophysical data for hydrogeological characterization. These advances include improved digital technology for data acquisition, processing and visualization, and improved geophysical technologies for near surface exploration. As described by Hubbard & Rubin (2005), hydrogeophysical approaches have been used for hydrological mapping (such as hydrostratigraphy or the depth to the water table and bedrock), estimation of hydrogeological properties (water content, porosity, hydraulic conductivity, lithology, water salinity) and hydrological monitoring (water infiltration, system responses to contaminant remediation).

However, there are still many obstacles that hinder the routine use of geophysics for hydrogeological characterization (Hubbard & Rubin, 2005). Perhaps the largest obstacle associated with the application of geophysics is the lack of systematic approaches for data inversion or integration, because to perform integration we must confront issues of scale, non-uniqueness, and uncertainty. Many different integration and estimation approaches have been used to fuse hydrogeological and geophysical data. The choice of approach typically depends on data density, project objectives, and the discipline and training of the interpreter. The second major obstacle is the need for petrophysical relationships to link hydrogeological parameters (such as porosity and type and amount of pore fluid) within typical near-surface environments to geophysical attributes (such as seismic velocity, electrical resistivity, and dielectric constant). Although substantial research has been devoted to exploring petrophysical relationships for consolidated formations common to petroleum reservoirs and mining sites, very little petrophysical research has been devoted to the less consolidated, low-pressure and low-temperature environments common to hydrogeological investigations. Challenges are also associated with the need to improve our understanding of less conventional geophysical approaches, and to develop a better assessment of the accuracy of geophysical data inversion approaches in the presence of noise, acquisition and inversion artefacts, and in real three-dimensional natural systems. Hydrogeophysical and tracer techniques enable hydrological properties and processes to be estimated over larger spatial scales and in a less invasive manner than conventional approaches. Although there are still many challenges associated with the use of geophysical methods for hydrogeological characterization, there is now a great body of evidence that illustrates the value of these data for understanding hydrological systems and improving the management of natural systems.

4.9.3 Future of remote sensing

Traditionally, field measures of hydrological properties and fluxes have tended to be point measures. Recent advances in remote sensing and data availability are also providing new insights into a number of hydrological problems and spatial data are increasingly being used in conjunction with hydrological models. The incorporation of spatial data is not however without complications. Remote sensing in hydrology is used for the identification and monitoring of relevant features such as snow-covered areas or evaporation. Depending on the item of interest, processing to extract that feature requires varying levels of interpretation, classification and validation. Traditionally, hydrological science has not been the primary focus in the development of remote sensors and platforms, although this has changed with the advent of recent and planned missions. Consequently, our current ability to exploit remote sensing as a tool for hydrological science is sensor- and platform-dependent. We are restricted by the sensor's spectral resolution, and monitoring wavelengths as well as the platform's sampling frequency and spatial resolution. In general, the parameters in hydrological models that are most commonly characterized by remote sensing are land-cover and snow-cover extent. An emerging technique is the use of variations of the gravity field of the Earth which can potentially be used to infer changes in hydrological storage.

Typical sensors and platforms that have been successfully used to extract vital hydrologic information include: Landsat TM, MSS, NOAA, SPOT and ERS-2 (land cover/use and vegetation); ERS-1 and -2, Radarsat, JERS-1 and SSM/I (soil moisture); SPOT ERS-1 and -2, Radarsat, Landsat TM and MSS (surface water); Nimbus 5 (spring runoff); ERS-1 and JERS-1 (temporal changes in snowmelt and soil moisture); Landsat and Topex-Poseidon (water depth); NOAA, SPOT, Landsat and MODIS (snow-covered area); GOES and Nimbus 7 (snow depth); DMSP, SSMI/I, MOS-1 and MSR (snow water equivalent); ERS-1 and -2 and Radarsat (wet snow); NOAA and Landsat TM (surface temperature); Meteor (solar radiation); AVHRR, ERS-2 and GOES (surface albedo); NOAA (evapotranspiration); CHARM and GRACE (gravity field, hydrological storage changes); Meteosat, GOES, SSMI/I and TRMM (precipitation).

Nearly all the development and testing of algorithms relies on ground-based measurements that are generally made at scales of less than a kilometre. Measurement length scales are surface- and time-dependent. In addition, many hydrological models are not designed to use the type of spatial data provided by remote sensing and the accuracy of variables/parameters derived from remotely-sensed data may be insufficient to warrant the development of new models. Data fusion, which involves combining remotely-sensed data with other information, such as digital terrain models, seems to offer the most promising approach for maximizing the availability of information at appropriate time and space scales for hydrological modelling.

Remote sensing will be an integral part of the path to understanding in hydrological science, supporting modelling and prediction, and assisting in the implementation of water resources policy. When hydrological activities that can benefit from remote sensing start to use it *operationally* (the application must produce an output on a regular basis or the remote sensing approach must be used regularly and on a continuing basis as part of a procedure to solve a problem; Rango, 1994; Rango & Shalaby, 1998), then its full potential will be realized. The role afforded by remote sensing to a particular hydrological activity will vary from minimal to critical. Groundwater studies may have only minimal use for remotely-sensed data but sea-ice monitoring and modelling is nearly 100% dependent on remote sensing, and hence cryosphere scientists will want to direct greater effort and interest to further researching this area in the next two decades.

While the operational application of remote sensing to many subdisciplines of hydrology may not be realized for decades, the issues plaguing operational applications are in fact very similar to those affecting research-directed and intermittent (i.e. nonoperational) applications. Before operational hydrology can become commonplace, several issues, discussed below, need to be addressed.

Financial Research into the integration and appropriate application of remote sensing in hydrology is no exception to the common plea for greater funding. But, more importantly, the relative expense of remotely-sensed data often prohibits its use in research, particularly in operational applications. Institutions and communities must find ways to decrease the costs of this data.

Revisiting old but persistent problems: atmospheric correction While applications using microwave technology are becoming more common, data collected at optical, near, middle and thermal infrared wavelengths will always be a staple of remotely-sensed applications to hydrology (particularly in view of the need to merge data collected at various wavelengths). This means that atmospheric contamination will continue to be an issue in the pre-processing stage. Considered as noise and a rather unglamorous problem, the removal of atmospheric contamination from reflectances and brightnesses needs to be revisited to provide more satisfactory solutions. Involving atmospheric scientists may help to progress this area of research and make atmospheric correction less complex and automatic. Data pre-processing, which is an essential part of remotely-sensed applications, should be automated and funding to develop such automation should be considered important.

Infrastructure, training and management Sufficient capacity in infrastructure, trained personnel, and data management and decision making is critical to the wide-spread operational use of remote sensing in hydrology. The importance of this support system cannot be understated and it should be developed or considered in tandem when determining data needs. Technical issues will develop related to data storage, processing, transferral and archiving; thus soliciting the assistance of those with expertise in data management is vital. In addition, improved communication is required between the different stakeholders involved so that decision makers understand the important factors affecting the availability, accuracy and application of remotely-sensed data and users communicate their data requirements to those in data management positions.

Appropriate sensor and platform designs For many applications, limitations exist due to coarse spatial and temporal resolutions and the inappropriate use of data in applications for which they were never intended (perhaps because of the wrong scale of application or spectral bandwidth). A wide variety of applications need better and finer spatial resolutions in panochromatic and multispectral imaging for all wavelengths. Greater options on polarization for SAR imagery would be ideal, as many sensors are only configured with single polarizations. Future payloads will help to fill the gap in large-scale hydrological data collection and give hydrologists greater access to data specifically designed for collecting various hydrological parameters at *specific scales*. Of course the full benefits of these data will only be realized with appropriate benchmarking and evaluation efforts.

Evaluation methods and benchmarking The development and evaluation of tools using remotely-sensed data will continue to rely on ground-truth data, perhaps even in an operational setting. *Benchmarking*, or algorithm inter-comparison projects, is an essential exercise in evaluation. It involves a variety of datasets to which different algorithms are applied, tested and their performance compared. Providing these kinds of evaluation datasets to test current and newly proposed algorithms for converting remotely-sensed data to a hydrological parameter will help assess the temporal and spatial universality of these algorithms (issues that have plagued modern algorithm use). Developing standard methods of evaluation is important for proper assessment, as well as knowing the accuracy of the ground-truth data used to assess the remotely-sensed data and the accompanying algorithm. One of the major difficulties with the use of ground-based measurements to develop and test algorithms from remotely-sensed data is the frequent incompatibility in scale between the datasets. Current validation efforts still rely on scaling from a point ground measurement to grid sizes as large as 1 km × 1 km (Stewart *et al.*, 1998). In order to use the multi-scale approach required in data fusion methods (suggested below), we must develop ground-based instruments that make measurements at scales compatible with scales of remotely-sensed data. Furthermore, when remotely-sensed data is used in hydrological models, the error in the data should be propagated to model output and reported accordingly.

<u>Analysis trends for the future: data fusion</u> Data fusion, integration and data mining are the future of remote sensing for hydrological purposes. Finding the most appropriate datasets and *fusing* them may lead to significant advances in hydrological science. Data fusion, in which data at different spatial, spectral or temporal scales are combined, has been identified as a means to solving numerous problems that have impeded the use of remote sensing in hydrology. The inherent difficulty in data fusion arises out of the very thing that makes it so attractive: the data are collected at different scales. Consequently, the scale of the original data must be respected, appropriate scaling techniques are required for multi-scale analysis and we must attempt to understand the processes occurring at different scales.

The importance of the above issues will receive greater recognition in new initiatives that seek to merge datasets to support long-term studies of global change, and emergency response and disaster management. Such initiatives include the Global Earth Observation System of Systems (GEOSS) which will attempt to unite observation and data collection systems around the globe (Lubick, 2005), and the Integrated Global Observing Strategy (IGOS) which seeks to provide a comprehensive framework for earth observation that will build upon the strategies of existing international space-based and *in situ* global observing programmes. These types of initiatives, that will lead the way in combining data, require a foundation in data fusion, mining and integration and research in this area should be at the forefront of research related to remote sensing applications in hydrology.

4.9.4 Future of water quality monitoring

By 2020 there will be an explosion in the quantity, nature, and temporal and spatial coverage of water quality data available to hydrologists. The development of micro-sensors (including MEMS - “labs on a chip” - and motes) will allow cheap, continuous, real-time measurements of physical and chemical components of water quality to be made in all components of the hydrological cycle (precipitation, soil water, plant water, surface water, groundwater). Parallel technical developments in communications, computing, electronics and e-science are expected to enable the real time acquisition, assimilation and analysis of these water quality data. Quality control of the vast streams of water quality data could be achieved automatically through models incorporated within the data processing phase.

Coupled with improved spatial datasets that are continuously updated from remote sensing (such as precipitation coverage, land use, vegetation cover, soil moisture content), these water quality datasets would enable an improved understanding of the relationships between climate, land use and management and water quality. Diffuse pollution of surface and groundwater resources could be detected in real time, allowing remedial actions to be taken immediately and thereby providing an improved understanding of the causes and effective ways of reducing and managing water pollution. In addition to the vast amounts of data available on the physical and chemical components of water quality, it is anticipated that hydrologists will also utilize other more holistic measures of water quality, such as micro-organisms, DNA signatures and biological indicators, in field investigations and modelling.

An interesting analogy with the current state of technology and modelling in water quality is the "carbon" scientific community in the mid-1990s. At this time the eddy covariance method of continuously assessing the carbon uptake of ecosystems in real time had just been developed and measurements were only being conducted at a few sites. Today (in 2006) there are carbon flux networks worldwide and data assimilation and modelling approaches are under development to process the vast quantities of data produced by such networks (e.g. Papale & Valentini, 2003; Williams *et al.*, 2005). Examples of the outputs that could be realized from such developments include the real time measurement of carbon sequestration to monitor the effectiveness of the Kyoto Protocol for stabilizing the carbon dioxide concentration of the troposphere. By building upon the experiences of the "carbon" scientific community, huge steps forward in the availability of low cost water quality data are anticipated by 2020 that will enable substantial advances to be made in our understanding of water quality relationships in ungauged catchments. Nevertheless, in promoting the development of new technologies within hydrology, we should be mindful of attendant cultural, political, social and environmental concerns, for example, access to data and environmental issues relating to the deployment of MEMS or motes.

Future water quality research, perhaps more than any other area of hydrology, will be driven by specific problems within individual catchments that require unique field programmes and models. Nevertheless it is anticipated that more generic water quality research will be conducted in a global network of experimental catchments with nested sampling programmes in order to fully understand climate, land use and water quality relationships. The key difference between the experimental catchments of the future and historical catchment studies is that people will be fully integrated into the field sites and models in the future, rather than excluded.

By 2020 hydrologists will be working seamlessly alongside scientists from other disciplines, particularly biologists, ecologists, chemists, electronic engineers and computer scientists, to develop new tracers, new analytical methods, better indirect indicators of water quality, new sensors and new models for processing the vast quantities of data generated and to yield an improved understanding of the controls on water quality.

4.10 DATA COLLECTION SYSTEMS AND STANDARDS

Rodda (1997) suggested that high quality, long-term data necessary for detection of global change remains a dream for many hydrologists and that rigorous standards for accuracy in data collection are lacking. While many instrumented field sites exist in various locations around the globe, changes in instrumentation, environment or the convenience of collecting the data, has resulted in the development of datasets that are less than optimal.

Standards for data collection need to be developed in order to provide information on water resources, whether at a global scale such as WHYCOS, the WMO's World Hydrological Cycle Observing System, or at the hillslope scale. The data collected at these "hydrological observatories" should be provided to those who can most benefit from this data. Aside from the development of the technology required to effectively measure the objective parameter/variable at the appropriate scale, issues that should be addressed in establishing a suitable database include:

- developing standards for accuracy and precision in data collection (data quality);
- developing standards for error and data quality reporting;
- developing quality assurance procedures;
- reporting on instrument change and malfunction;
- selecting the appropriate database or Geographic Information System, the appropriate data model, and metadata standards;
- establishing trained personnel assigned to update the database.

4.11 CONCLUDING REMARKS

This chapter has examined some emerging technologies for quantifying essential hydrological fluxes that can lead to our understanding and improved prediction of hydrological processes. The availability and quality of hydrological data will drive much of the model development that is the subject of the next chapter. A great deal of progress is needed to improve our data collection activities and this will in turn help to advance hydrological modelling. Hence, raising the status of data collection and management activities in hydrological research is essential. "Management" can also refer to research efforts that look solely at the feasibility of merging various hydrological data and determining the appropriateness of data fusion techniques, although the latter is plagued by issues of scale.

To understand hydrological science requires monitoring processes that operate at a variety of scales. We need to address scaling issues and to be aware of two complementary concepts: the *domain of scale* and the *scale threshold*. The former corresponds to the regions of the scale spectrum over which, for a particular phenomenon, patterns do not change or change monotonically with changes in scale. These domains are separated by *scale thresholds* that are relatively sharp transitions where a shift in the relative importance of variables influencing a process occurs (Marceau, 1999). Identifying these domains and thresholds, and, eventually developing the scaling laws required to move between domains, requires sufficient and intensive monitoring at a variety of scales. Marceau (1999) has identified hydrology as one field which has made progress on the scale issue through the use of *physically-based distributed models* which are tested at different scales to identify the dominance of certain parameters at various scales. Nevertheless, Gupta & Sorooshian (1997) stated that the hydrological community needs to improve communications between modellers

Fig. 4.5 *Prototype of the hydrologist of the future; hydrologists need to increase the integration of modelling and field measurement activities.*

and data collectors, and that intercomparison of models can also help to identify model errors. The next chapter will discuss these issues and examine the necessary but dichotomous relationship between models and data in hydrology. A cartoon of the hydrologist of the future is shown in Fig. 4.5, illustrating the need for increased integration between hydrological measurement and modelling with the ultimate aim of addressing real water resource management issues.