Contributions of groundwater conditions to soil and water salinization

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Abstract Salinization is the process whereby the concentration of dissolved salts in water and soil is increased due to natural or human-induced processes. Water is lost through one or any combination of four main mechanisms: evaporation, evapotranspiration, hydrolysis, and leakage between aquifers. Salinity increases from catchment divides to the valley floors and in the direction of groundwater flow. Salinization is explained by two main chemical models developed by the authors: weathering and deposition. These models are in agreement with the weathering and depositional geological processes that have formed soils and overburden in the catchments. Five soil-change processes in arid and semi-arid climates are associated with waterlogging and water. In all represented cases, groundwater is the main geological agent for transmitting, accumulating, and discharging salt. At a small catchment scale in South and Western Australia, water is lost through evapotranspiration and hydrolysis. Saline groundwater flows along the beds of the streams and is accumulated in paleochannels, which act as a salt repository, and finally discharges in lakes, where most of the saline groundwater is concentrated. In the hummocky terrains of the Northern Great Plains Region, Canada and USA, the localized recharge and discharge scenarios cause salinization to occur mainly in depressions, in conjunction with the formation of saline soils and seepages. On a regional scale within closed basins, this process can create playas or saline lakes. In the continental aquifers of the rift basins of Sudan, salinity increases along the groundwater flow path and forms a saline zone at the distal end. The saline zone in each rift forms a closed ridge, which coincides with the closed trough of the groundwater-level map. The saline body or bodies were formed by evaporation coupled with alkaline-earth carbonate precipitation and dissolution of capillary salts.

Résumé La salinisation est le processus par lequel la concentration des sels dissous dans l'eau et les sols s'accroît sous l'effet de processus naturels ou anthropiques. L'eau est perdue par l'une ou l'autre combinaison de quatre principaux mécanismes : l'évaporation, l'évapotranspiration, l'hydrolyse et la drainance entre aquifères. La salinité augmente depuis les limites des bassins jusqu'au fond des vallées et le long des axes d'écoulement souterrain. La salinisation est expliquée au moyen de deux principaux modèles chimiques développés par les auteurs : l'altération et le dépôt. Ces modèles sont en accord avec les processus géologiques d'altération et de dépôt qui ont formé les sols et qui recouvrent les bassins versants. Cinq processus d'évolution de sols sous climats aride et semi-aride sont associés à l'eau et à des formations aquifères. Dans tous les cas présentés, l'eau souterraine est le principal agent géologique qui transporte, accumule et dépose les sels. A l'échelle de petits bassins versants dans le sud et dans l'ouest de l'Australie, l'eau est consommée par évapotranspiration et par hydrolyse. L'eau souterraine salée coule le long des berges de rivières et s'accumule dans des paléochenaux, qui fonctionnent comme des zones de stockage de sels, et finalement s'écoule dans des lacs, où la plupart des eaux salées se concentrent. Dans les formations bosselées des grandes plaines du nord du Canada et des États-Unis, des scénarios d'alimentation et de décharge localisées conduisent la salinisation à se produire surtout dans les dépressions, en association avec la formation de sols et d'infiltrations salins. A l'échelle régionale dans les bassins fermés, ce processus peut être à l'origine de playas ou de lacs salés. Dans les aquifères continentaux des bassins de rift du Soudan, la salinité augmente le long des axes d'écoulement souterrain et forme ainsi une zone saline à leur extrémité. La zone saline de chaque rift constitue une crête, qui coïncide avec le creux piézométrique. Le ou les ensembles salins se sont formés par évaporation couplée à la

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Supplemental material The colour version of *Figure 3* has been deposited in electronic form and can be obtained from http://link.springer.de/link/service/journals/10040

précipitation d'un carbonate alcalino-terreux et à la dissolution de sels par capillarité.

Resumen La salinización es el proceso por el cual la concentración de sales disueltas en el agua o el suelo aumenta por causas naturales o antrópicas. El contenido de agua se reduce por uno o una combinación de los siguientes mecanismos: evaporación, evapotranspiración, hidrólisis y goteo entre acuíferos. La salinidad aumenta en la dirección del flujo subterráneo, desde los puntos de recarga hacia los de descarga. La salinización se puede explicar por dos modelos químicos que se pueden definir como erosión y deposición y que están de acuerdo con los clásicos procesos geológicos que tienen los mismos nombres. Se conocen cinco procesos de salinización de suelos ligados al agua o a la extracción de la misma en climas áridos o semiáridos. En todos los casos que se presentan el agua subterránea es el principal agente geológico de transmisión, acumulación y descarga de sales. A escala de pequeña cuenca, en Australia Occidental y del Sur, el agua se pierde mediante evapotranspiración e hidrólisis. El agua subterránea salina fluye a lo largo de los cauces de los arroyos y se acumula en los paleocanales, que actúan como depósitos, hasta que finalmente descarga en los lagos, donde se acumula la mayor parte de la salinidad. En los hummocks de la Zona Norte de los Grandes Llanos de Canadá y EEUU, la recarga localizada y el tipo de descarga provocan que la salinización tenga lugar fundamentalmente en las depresiones. A escala regional y en cuencas cerradas este mismo mecanismo provoca playas o lagos salinos. En los acuíferos continentales de los rift del Sudán, la salinidad se incrementa a lo largo de las líneas de flujo subterráneo y se concentra en el límite distal. La zona salina en cada rift forma una cresta cerrada que se manifiesta en los mapas piezométricos. El cuerpo o cuerpos salinos se formaron por evaporación, acoplada con la precipitación de carbonatos y la disolución de sales capilares.

Key words geomorphology · groundwater flow · salinization · chemical models · geological agent

Introduction

Salinization of land and water is brought about by physical and chemical processes that increase concentrations of salt in soil and water. The processes responsible for the development of saline land and water are complex and intimately related to the transport of dissolved mass in groundwater flow systems. The redistribution of soluble salts accumulated in a catchment is evident mainly in topographically lower areas by terminal salt lakes, dry salinas, and areas of saline seeps and scalds.

The countries affected by salinization are mainly located in arid and semi-arid regions and include areas in North and South America, Australia, China, Commonwealth of Independent States, India, regions in the Mediterranean and Middle East, and Southeast Asia. The world loses about 10 ha of arable land every minute, 3 ha of which are lost by salinization (Ghassemi et al. 1995). Secondary salinization induced by human activity occurs in irrigated and non-irrigated dryland salinity areas.

This paper discusses the processes of soil and water salinization as they occur in different geological, hydrogeomorphological, agricultural, and climatic settings. Specifically, the objectives are to characterize and describe the groundwater conditions contributing to existing salinity problems. Examples from open and closed systems from central Africa (Sudan), North America, and South and Western Australia are given to illustrate the different settings, processes and modes of development of salinity. These are summarized in *Table 1* and general locations are given in *Figure 1*.

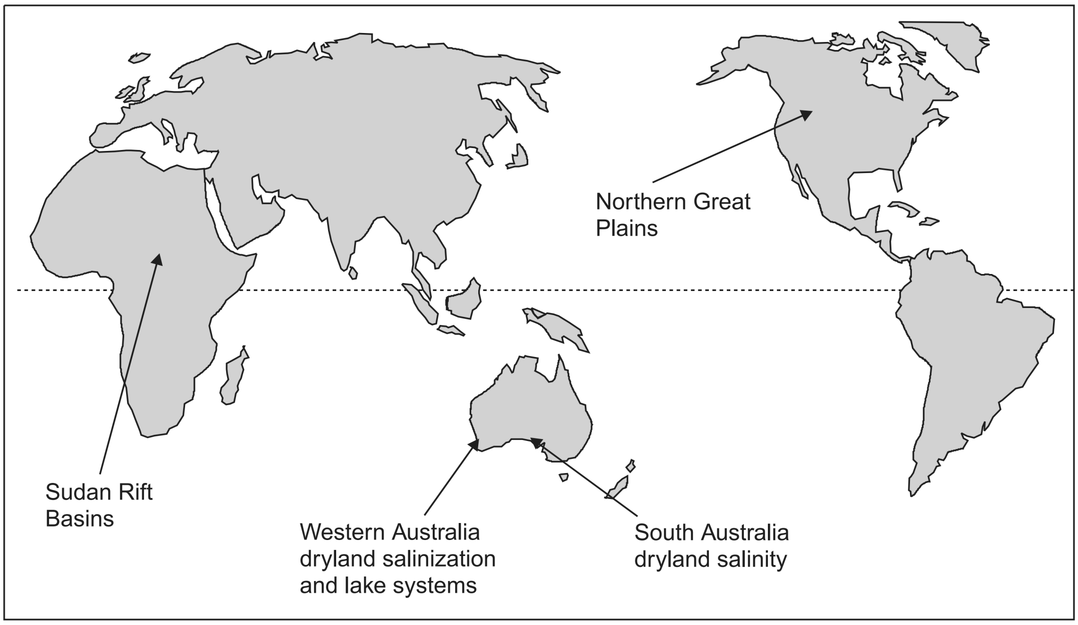
Throughout this paper, the following four mechanism of water uptake and salt concentration are recog-

Table 1 Case studies presented in the paper

System	Expression	Chemistry	Control[1]	Flow system	Locality/Country
Open	Discharge areas	SO_4, NaCl	Hydromorphic	Intermediate	South Australia
Open	Discharge areas	NaCl	Hydrogeomorphic	Local, intermediate, and regional	Wheat belt – Western Australia
Open	Series of lakes	NaCl	Hydrogeomorphic	Regional	Salt River System – Australia
Closed	Saline lake	NaCl	Hydrological and Hydrogeological	Intermediate	Lake Deborah – Western Australia
Closed	Rift basin	Ca-$MgCO_3$–SO_4–Cl	Hydrogeological	Regional	Buried lakes – Sudan
Closed	Discharge areas	Na-$MgSO_4$	Hydrogeomorphic	Local–intermediate	Northern Great Plains of America

[1] Hydromorphic: soils, surface, and groundwater interaction. Hydrogeomorphic: soils, geology, surface, and groundwater interaction. Hydrological: surface water. Hydrogeological: groundwater

Figure 1 Locations of case studies in the Northern Great plains (USA and Canada), South Australia and Western Australia, and Sudan Rift Basins (Africa)



nized: evaporation, evapotranspiration, hydrolysis, and leakage. Salt accumulates when mineralized groundwater at or near the ground surface continually evaporates and causes minerals to precipitate; by evapotranspiration where infiltrating recharge water is continually taken up by plants and salt is concentrated in the unsaturated zone; by hydrolysis where water is taken up in the formation of new minerals in the weathering processes; and by leakage between aquifers through confining beds. First-, second-, and third-order catchments refer to catchments with first-, second-, and third-order streams, respectively.

Hydrogeomorphic Controls on Groundwater Flow

The distribution of hydrogeomorphic units in a catchment is controlled by the geologic formations from which the units developed (Salama et al. 1994). Each characteristic geological formation leads to the development of a particular unit. Groundwater in each unit is controlled by its geomorphic characteristics (topography and hydrostratigraphy). Groundwater levels are deep in undulating upland areas and near the surface in topographic lows of the landscape. In general, the water-table configuration is a replica of the topography. However, in arid and semi-arid climates the outline of the water table is subdued, and the hydraulic gradients are generally less steep. Slope, break of slope, and curvature control where groundwater discharge takes place. Topography generates groundwater flow systems of different orders that correspond to the dimensions of the relief of a catchment's surface (Tóth 1963). Three orders of flow systems may be distinguished, namely, local, intermediate, and regional. Each flow system has a recharge, transfer (midline), and discharge area.

Groundwater Control on Salinization in Arid/Semiarid Areas

Mechanisms

The spatial distribution of saline land and water in a catchment is related to the hydrogeomorphology (e.g., topography and hydrostratigraphy) and associated groundwater flow systems. The physical and chemical processes responsible for the development of saline soils are shown in *Figure 2*. They involve the mineralization of the groundwater, the physical transport of dissolved salts, the discharge of saline baseflow into streams and lakes, and the precipitation of salts within the soil zone. Most of the salt in the groundwater system comes from input loading, which includes aerosol salts, salt dissolved in the water recharging the system, and salt contributed from mineral dissolution within the groundwater flow system. The most important process that adds salt to groundwater is mineral-dissolution reactions in the subsoil and, to a lesser extent, along the entire flow system.

Advection and dispersion are the two physical-transport processes responsible for the transport of

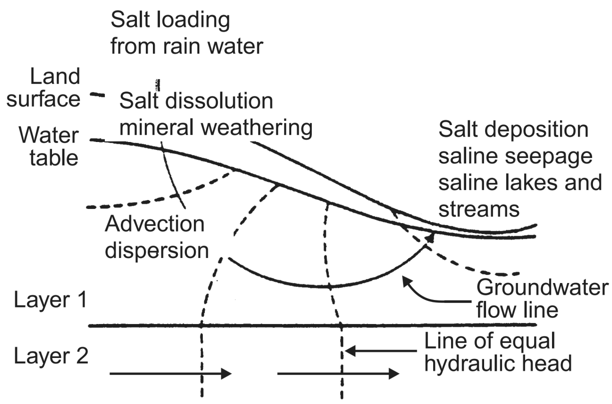


Figure 2 Conceptual model of soil and water salinization

dissolved salts from recharge to discharge areas. The pattern of transport by advection is influenced by the direction and rate of groundwater flow. Of importance in advection is the ability of the flow systems to collect excess recharge and groundwater flow over relatively large areas, and to focus that flow into localized discharge areas. The distance between the recharge and discharge areas is in general a few kilometers or less. Groundwater flow is mainly laterally downslope and occurs most often over shallow, less permeable bedrock.

Transport by dispersion is most important at small scales (<1.0 m) as far as the generation of saline soils is concerned. At a larger scale (>1.0 m), dispersion occurs as a consequence of porous-medium heterogeneity, which is responsible for changes in direction and rate of groundwater flow. The effect of dispersion at this scale is to spread and homogenize mass in the system and to attenuate high salinity.

Soil salinization occurs in areas of groundwater discharge or a rising water table when mineralized porewater at or near the ground surface continually evaporates and causes minerals to precipitate. A key factor controlling the amount of evaporation is the depth of the water table below the surface. In general, evaporation is minimal when the water table is below 1.5–3.0 m depth. Rising water tables can also lead to the formation of saline soils in recharge areas. However, groundwater discharge areas are commonly sites of the most active soil salinization, because salt fluxes are greatest there.

Sources of excess recharge

The above factors are prerequisite for the development of land and water salinization. However, it is the "excess" net recharge that triggers salinity formation in arid/semi-arid areas. For example, in the southern part of the Australian continent, saline soils and seeps that existed prior to settlement were the result of additional moisture formed by natural conditions, such as ponding and above-average annual rainfall. Secondary salinization is caused by man-induced actions, such as summer fallow, irrigation, dam construction, and clearing of native vegetation. Dryland salinity in a catchment is the hydrogeological response to the clearing in the landscape and the replacement of native vegetation with shallow-rooted crops. Groundwater recharge increases and water tables rise or the pressures in confined aquifers increase, causing an upward leakage to the phreatic aquifers. The additional water available on catchments converted to dryland farming is estimated to be 20–100 mm/a (Holmes and Talsma 1981).

Sources of salt

Salts present in the subsoil or deeper formations may have accumulated from ocean salt carried in rain (aerosols) into a previously highly leached landscape (e.g., Western Australia); or the salts are derived from weathering of the mineral constituents in the soil or may be present initially in strata of marine origin (e.g., North American Great Plains). The geographic distribution of aerosols decreases with increasing distance from the coast. The accumulation of this salt over thousands of years is considerable, and the amount of salt stored in the soil profile depends mainly on the soil type and the mean annual rainfall. The salt content is generally higher in clayey soils than in sandy soils, and the salt content changes inversely with average annual rainfall. The dominant soluble salts in saline soils mostly comprise chlorides, sulfates, and bicarbonates of sodium, calcium, and magnesium.

Evolution of groundwater chemistry

The evolution of the groundwater chemistry and the evaporite mineralogy of saline seeps is closely related to the hydraulic regime of groundwater systems. Groundwater flow systems are continuous transport mechanisms having different directions and dynamic conditions in their hydraulic regime (Tóth 1984). These are summarized in *Table 2*.

Groundwater control on soil processes

Salt-affected soils occur under different environmental conditions and have diverse morphological, chemical, physical, physico-chemical, and biological properties. Szabolcs (1991) groups all salt-affected soils into one

Table 2 Groundwater chemistry and hydraulic regimes of flow systems. (After Tóth 1984)

Hydraulic regime	Conditions	Chemical evolution
Recharge area	Rain water, low in TDS, high CO_2, downward flow, cross–formational flow	Increase in TDS; HCO_3, SO_4, Ca, and Mg are dominant species; spatial variability of water chemistry as different lithologies are traversed
Midline area	Moderately charged recharge water, low CO_2, mainly lateral flow	TDS increases with length of flow and residence time; Na, Ca, Mg, HCO_3, SO_4, and Cl are dominant species
Discharge area	Mineralized water, upward and cross–formational flow, mixing with infiltrated fresher water	High TDS at land surface; SO_4, Cl, and high CO_2. Na is dominant species; precipitation of minerals

large group of soils because they have, as their common feature, the dominating influence of electrolytes on their formation and properties. Various soil sub-groupings are based on the types of electrolytes causing the salinity. The following properties are common for salt-affected soils in most soil-classification systems:

1. Morphology of the soil profile (presence or absence of diagnostic horizons)
2. Significant physical properties (mainly for the Solonetz group)
3. Chemical and physico-chemical properties:
   (a) Salt content, salt composition, and salt distribution in the profile and, in some cases, also in the groundwater
   (b) Exchangeable sodium percentage and sodium adsorption ratio
   (c) pH conditions and the existence of sodium carbonate

Each soil system of the soil landscape may be linked to a specific water-flow system (Fritsch et al. 1992). Because the hydrology of a catchment is in disequilibrium and the surface and groundwater systems are in an unsteady state, soil processes are activated that replace earlier soil processes, such as saprolitization and ferralitization. Soil systems change in nature and extent until they are in equilibrium with the water-flow system. When the soil features are in phase with the modern hydrological regimes, the whole ecosystem is in equilibrium.

Five soil-change processes are associated with waterlogging and water flow: (1) redoximorphism (oxidation/reduction); (2) eluviation and illuviation (clay movement); (3) salinization and solonization (salinity and sodicity); (4) sulfidization and sulfuricization (sulfidic conditions); and (5) water erosion. These processes lead to the formation of the different hydromorphic domains (the association of each soil landscape with a certain water-flow system). These five soil processes are more recent and, in Australia, are due to land clearing since European settlement. The resulting disequilbrium develops severe land degradation problems as a result of rising saline sulfatic groundwater tables and perched water tables, particularly when they act together in downslope positions of the landscape. Examples are shown in *Figure 3*.

Above seepage areas, redoximorphism (waterlogging) induces clay eluviation near the surface and salinization at depth. In seepage areas, waterlogging and salinization near the surface induce sulfidization, where sulfide minerals (e.g., pyrite, or FeS_2) form in present-day reducing environments. The black sulfidic conditions in soils result from the crystallization of secondary fine-grained pyrite framboids by the action of sulfur-reducing bacteria in the presence of organic matter. This in turn induces sheet erosion. When the water table rises to the surface, seepage areas are flushed by fresh water, and solonization and sulfuricization takes place; the latter is the process by which sulfides are transformed to sulfates, sulfuric acid, and accompanying minerals (e.g., jarosite) following oxidation (e.g., when drained or disturbed). These processes in turn induce sheet erosion and activate gully erosion.

Incision of seepage areas by deep gullies lowers the groundwater level and (1) restricts the development of redoximorphism, salinization, and sulfidization; and (2) favors the development of eluviation, solonization, and sulfuricization. Incision of seepage areas also leads to deposition of sediments in down-stream areas. These deposits tend to raise the groundwater level locally and re-initiate the processes of salinization and sulfidization. These soil processes are supportive, particularly in seepage areas, but soil erosion controls the development of the others. Effectively, waterlogging and dryland salinity will continually increase until deep erosion incisions occur in the landscape.

Case Study I. Groundwater Controls on Dryland-Salinity Development and Saline-Soil Formation in South Australia

Generally speaking, Australian soil landscapes are extremely variable and complex due, in part, to much of the continent's great age. In adjacent landscape positions, deeply weathered soils that contain ancient stored salt are juxtaposed with very youthful soils on partly weathered rocks that are generating salt as a result of contemporary weathering processes.

Salt-Affected Soils in Australia: Soil Salinization

On a continental scale, Australia has the highest proportion of salt-affected soils in relation to total surface area in the world (Szabolcs 1991). Sodic and saline soils occupy almost 2106 km^2 and 3910 km^2, respectively (Northcote and Skene 1972). The sodic:saline ratio of 5.17 is 4.4–10.3 times that reported for other continents, and is consistent with the high proportion of sodium present in soil solutions and groundwaters. In Australia, most sodium-affected soils are the result of past inundation by brackish water supplemented possibly by cyclic salt. The result is that in subsoils chloride is the dominant anion and exchangeable Mg/Ca ratios are high. In California, the dominant anion in sodium-affected soils is sulfate.

Effects of adsorbed sodium on clay dispersion are most pronounced in dense alkaline subsoils that constitute about 86% of Australian sodic soils (Northcote and Skene 1972). The impact of soil sodicity on the environment is an important land-degradation issue in Australia. Both primary and secondary sodification of soils can cause undesirable changes in soil structure, severe hillslope erosion, waterlogging, and erosion of downstream watercourses.

Causes and Processes of Dryland Salinization

Since European settlement in Australia, widespread replacement of deep-rooted perennial native vegeta-

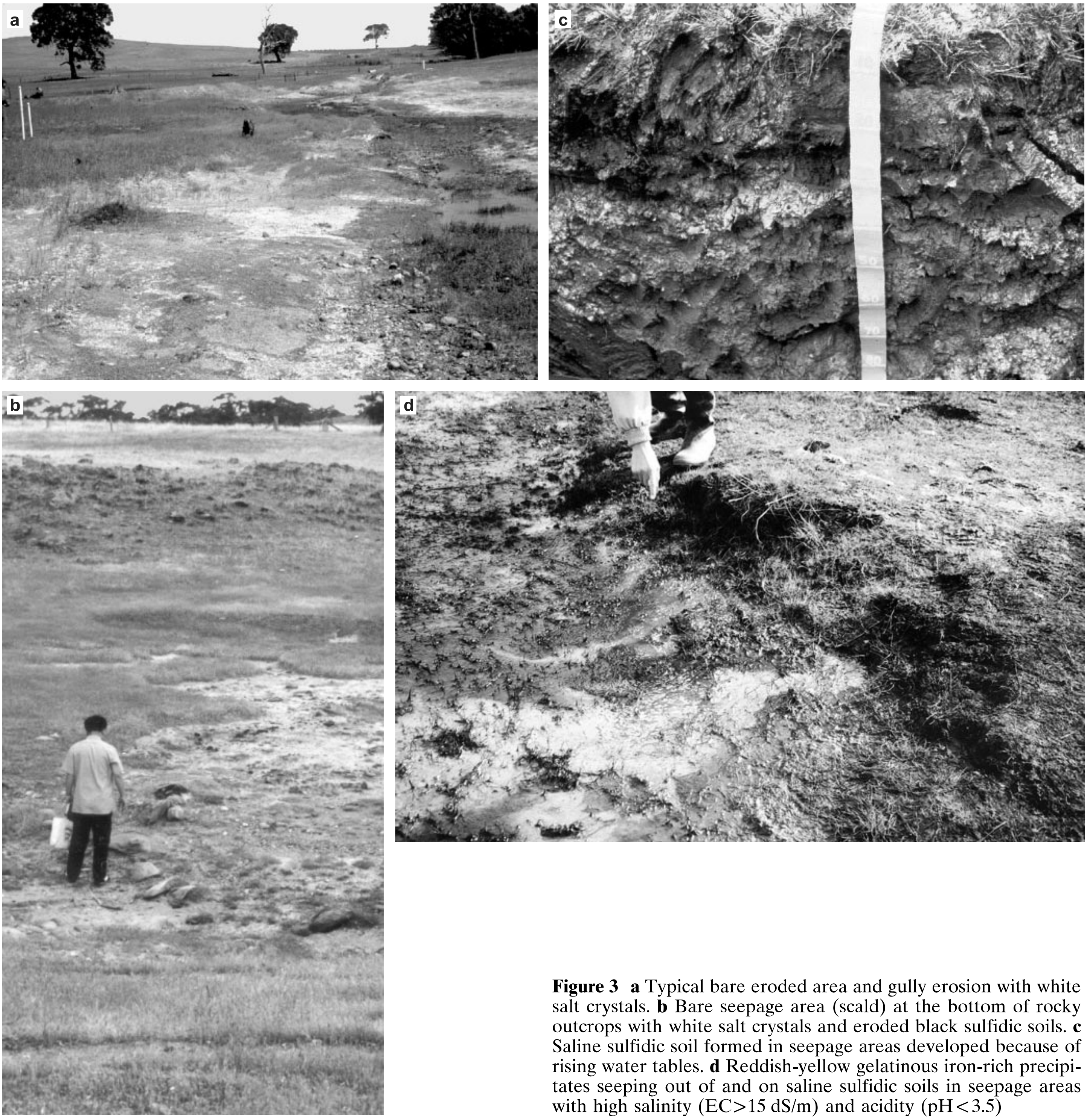


Figure 3 **a** Typical bare eroded area and gully erosion with white salt crystals. **b** Bare seepage area (scald) at the bottom of rocky outcrops with white salt crystals and eroded black sulfidic soils. **c** Saline sulfidic soil formed in seepage areas developed because of rising water tables. **d** Reddish-yellow gelatinous iron-rich precipitates seeping out of and on saline sulfidic soils in seepage areas with high salinity (EC>15 dS/m) and acidity (pH<3.5)

tion (often Eucalyptus species) with shallow-rooted annual plants, which not only use but also tend to intercept less water, has led to the development of dryland salinity (Peck and Williamson 1987). This deforestation has allowed more water to enter the groundwater system, causing water levels to rise and leading to mobilization of any stored salt both in and between catchments. Where water rises to within 1–2 m of the soil surface, and if there is salt mobilization, the soil becomes salt-affected as a result of evapotranspiration. A further problem associated with dryland salinization is that whereas the hydrological response to deforestation in some cases is rapid, elsewhere it is very slow, taking 50–100 yr for the effects to become evident.

Salt-affected soils in Australia, as defined by Northcote and Skene (1972) and presented here in simplified form, are extensive. However, the secondary saline soils that form following clearing (or irrigation) are more closely associated with the sodic soils. Work done in Western Australia (e.g., Bettenay et al. 1964) and Queensland (Gunn and Richardson 1979) show that the deeper subsoils in mottled and pallid zone materials contain substantial amounts of soluble salts, ranging in amounts from 10–100 kg m^{-2}. This accumulation of stored salt originates from several possible sources: the ocean via rainfall, weathering of soil and rock materials, and marine deposition in earlier geological periods (Isbell et al. 1983). Following the clearing of upland "lateritic" areas, saline seepages develop rapidly on slopes, particularly in certain higher rainfall areas of Victoria (Jenkin 1981) and Queensland (Gunn 1967).

Similarly, in the high rainfall zones of the Mount Lofty Ranges of South Australia, Fitzpatrick et al. (1996) have identified the rapid formation of saline seepages that contain high concentrations of mobile sulfate, iron, and manganese in surface and subsurface horizons. These soils are frequently waterlogged and are associated with or derived from particular geological formations that contain pyrite. Fluctuating redox conditions generated by waterlogging lead to formation of pseudo-acid sulfate soils (i.e., becoming strongly acidic with iron, manganese, and sulfate toxicities), which often contributes to structural degradation and erosion (Fitzpatrick et al. 1992).

Soil Salinization, Sodification, and Water Quality

The patterns of subsurface water movement are often very complex in Australian landscapes. Groundwater may occur as an unconfined aquifer in one part of a catchment leading into a confined aquifer farther down-slope due to the existence of layers (often paleochannels) with different hydraulic properties. Case studies conducted in the Mt. Lofty Ranges, South Australia by Fitzpatrick et al. (1992, 1996) illustrate the close and important inter-relationships between salinity and sodicity in the context of soil–water–landscape processes. Two types of sodic soils have been identified in Australia: naturally sodic soils, which are related to the parent material (Isbell et al. 1983); and secondary sodification processes arising from man-made activities. Both of these processes are influenced by climate, geology, topographic positions, soil, and cultural conditions, all of which determine the nature, extent, and severity of sodic soils. In South Australia, the distribution of sodic soils occurs largely in areas with rainfall exceeding 500 mm.

Recharge and throughflow on sloping lands and discharge at lower topographic positions also contribute to sodification and to lateral movement of colloids into streams. Such processes predominate in duplex soils, in which the dense sodic B horizon restricts downward movement of water. This constraint to infiltration leads to waterlogging, tunnel erosion, and enhanced lateral movement of water as surface runoff and as shallow groundwater flow in sloping land. Both waterlogging and loss of water through lateral flow influence crop productivity through fertility constraints and poor soil structure.

Land use has a profound effect on the quality of water draining duplex sodic soils. For example, in the Mt. Lofty Ranges, the sodicity [EC (Electrical Conductivity) and SAR (Sodium Adsorption Ratio)] of soil water draining duplex soils is also reflected in the amounts of dissolved organic carbon and the dispersed clay content of the stream waters. Dissolved organic carbon generated by sodicity enhances dispersion and, hence, mobilization of colloid material.

Groundwater Controls on Soil Formation and Salinization

In South Australia, piezometric data indicate that two distinct rising water-flow systems commonly occur in the catchments. A seasonal, fresh, perched water table develops between May and October on relatively impermeable subsoil layers. A rising water table occurs in the weathered and fractured rocks (Fitzpatrick et al. 1996). The groundwater is generally saline (EC 6–13 dS/m) and contains high concentrations of Mg, SO_4, and Fe. Preferentially flowing through old root channels and vertical cracks, the groundwater seeps to the surface under pressure in footslope positions. In winter months, some waterlogged soils in footslope or flat positions acquire distinctive reddish-yellow gelatinous iron-rich precipitates on their surfaces (*Figure 3*). Within the sub-surface layers of these soils, distinctive black blotches that are often quite smelly and soggy may be present due to the presence of sulfidic materials (*Figure 3*). In summer months, or when these soils are drained and are dry, reddish-yellow impermeable iron-rich crusts may occur on the soil surface (Fitzpatrick et al. 1996; *Figure 3*). Using x-ray diffraction and electron microscopy, each differently colored iron-rich feature was observed to contain certain specific types of iron minerals such as pyrite (black iron sulfide with a rotten egg smell), ferrihydrite (reddish brown), goethite (yellow), and schwertmannite (reddish-yellow).

Construction of Mechanistic Models of Landscape Evolution and Pedogenesis

Mechanistic models of landscape evolution and pedogenesis (e.g., Fritsch and Fitzpatrick 1994; Fitzpatrick et al. 1996) were constructed linking (or matching) the field and laboratory data directly to the various layers in the toposequence. The models clearly distinguish modern (since European land clearing, 130 yr ago) and relict (late Mesozoic) soil and hydrological processes. The relative abundance of each iron-rich feature and their distribution in relation to each other was used to establish a stepwise sequence or chronological order in the way each mineral formed and was transformed in these soils, and the degree of waterlogging that they represented (Fitzpatrick et al. 1992, 1996). Results of investigations in the Mt. Lofty Ranges and Dundas Tablelands in South Australia indicate that accumulation and oxidization of iron and sulfur in seasonally rising saline groundwater and surface water cause impermeable soil layers to form in water-discharge areas (Fitzpatrick et al. 1992, 1996). Research work has focused on the recent soil–landscape processes (namely, rising saline-sulfatic groundwaters associated with tree clearing since European settlement, 150 yr ago) that explain the complex transformation of fertile soils to saline sulfidic marsh soil. During the wet winter months, discharge areas creep up the slopes, thereby causing pristine duplex soils to be transformed into soggy saline and often sulfidic soils. In summer, unsightly eroded "iron ochre scalds" develop due to the oxidation of sulfur and iron to form sulfuric acid and various iron oxides (ferrihydrite, goethite, and schwertmannite), which causes soil pores to clog, especially in combination with sodic and finely dispersed clay particles.

Case II. Groundwater Controls on Salinity Development in the Wheat Belt of Western Australia

Geomorphology and Geology

The wheat belt of Western Australia, where most of the water and land salinization is taking place, is within the Western Gneiss Terrain and the Southern Cross Province of the Yilgarn Craton of Western Australia (Trendall 1990); locations are shown in *Figure 4*. The Western Gneiss Terrain is formed mainly of deformed, metamorphosed banded gneiss and intrusions of granite, whereas the Southern Cross Province is primarily composed of greenstones, gneisses, and granitoids (Griffin 1990). The Craton is cut by widespread dolerite dykes and is covered by transported or residual, unconsolidated to indurated regolith, underlain by a sequence of weathered and fractured bedrock material.

Catchments are usually bordered by lateritic duricrust and granitic outcrops. Lineaments, in the form of dolerite dykes with associated basement highs, occur in most catchments. Many of these, and also several outcrops, cut across drainage lines. The present drainage systems are heavily influenced by the structure within the underlying bedrock (Salama and Hawkes 1993). Another feature common in these catchments is the presence of an intricate system of relict channels, particularly in the valley floors.

Three aquifer systems are present in these catchments: (1) an unconfined aquifer is present above a clay layer in the sandplain and valleys; (2) a semi-confined aquifer occurs below this clay layer; and (3) a deeper semi-confined to confined regional aquifer occurs in the weathered zone above the bedrock. The deeper semi-confined to confined regional aquifer also extends into the sediments of the relict channels present in the valleys and the flanking areas in the Wallatin Creek catchment.

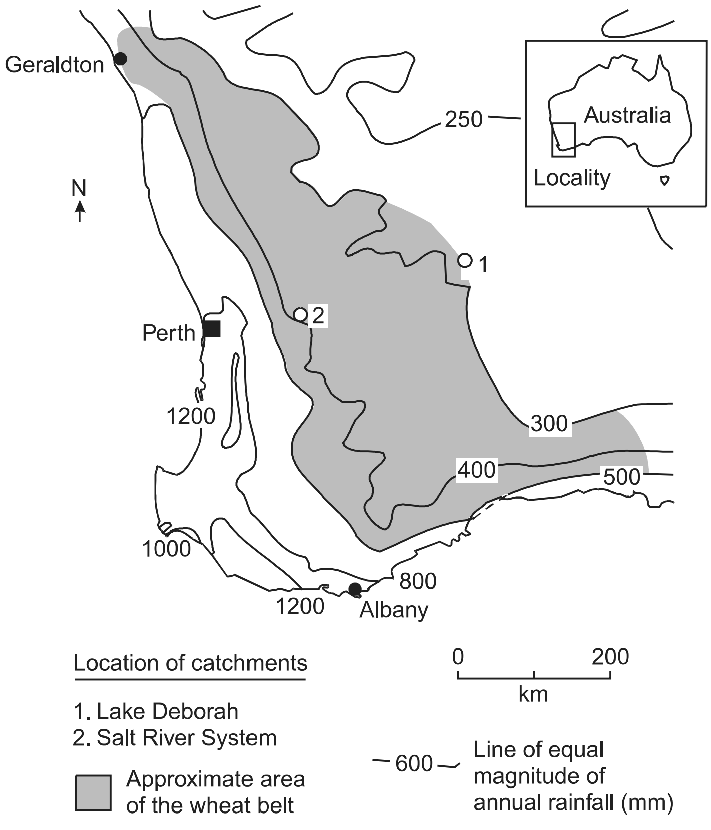


Figure 4 Locations of the wheat belt, Lake Deborah, and the Salt River system in Western Australia

Groundwater-Level Fluctuations and Groundwater Flow

Fluctuations in groundwater level and pressure indicate changes in groundwater behavior. Water-level changes can be used to distinguish aquifer types by their responses to atmospheric pressure changes (Jacob 1940; Gilliand 1969). Streams become effluent or influent, depending on the elevation of groundwater levels in relation to the base level of the stream. Regional groundwater flow patterns can also be controlled by the water-table configuration (Tóth 1962, 1963; Salama et al. 1993a). Local variations in the water table control the shallow and local flow system (Tóth 1963), and changes in groundwater levels enable changes in groundwater storage to be evaluated and areas of recharge and discharge to be delineated (Salama et al. 1993e).

Four patterns of water-level fluctuations (Salama et al. 1993e) are used in this work to classify the rate of rise and decline in water levels; these patterns are illustrated in *Figure 5*.

1. Monotonically rising water levels (*Figure 5a*), which occur in the upper reaches of catchments where the water levels are relatively deep (>5 m from the surface). The rate of rise is 0.5–2.0 mm d^{-1}.
2. Continuously rising water levels (*Figure 5b*), which occur in wells in mid-slope areas of catchments. The water levels exhibit a sinusoidal pattern with a rising trend. The rate of rise is 0.1–1.0 mm d^{-1}.

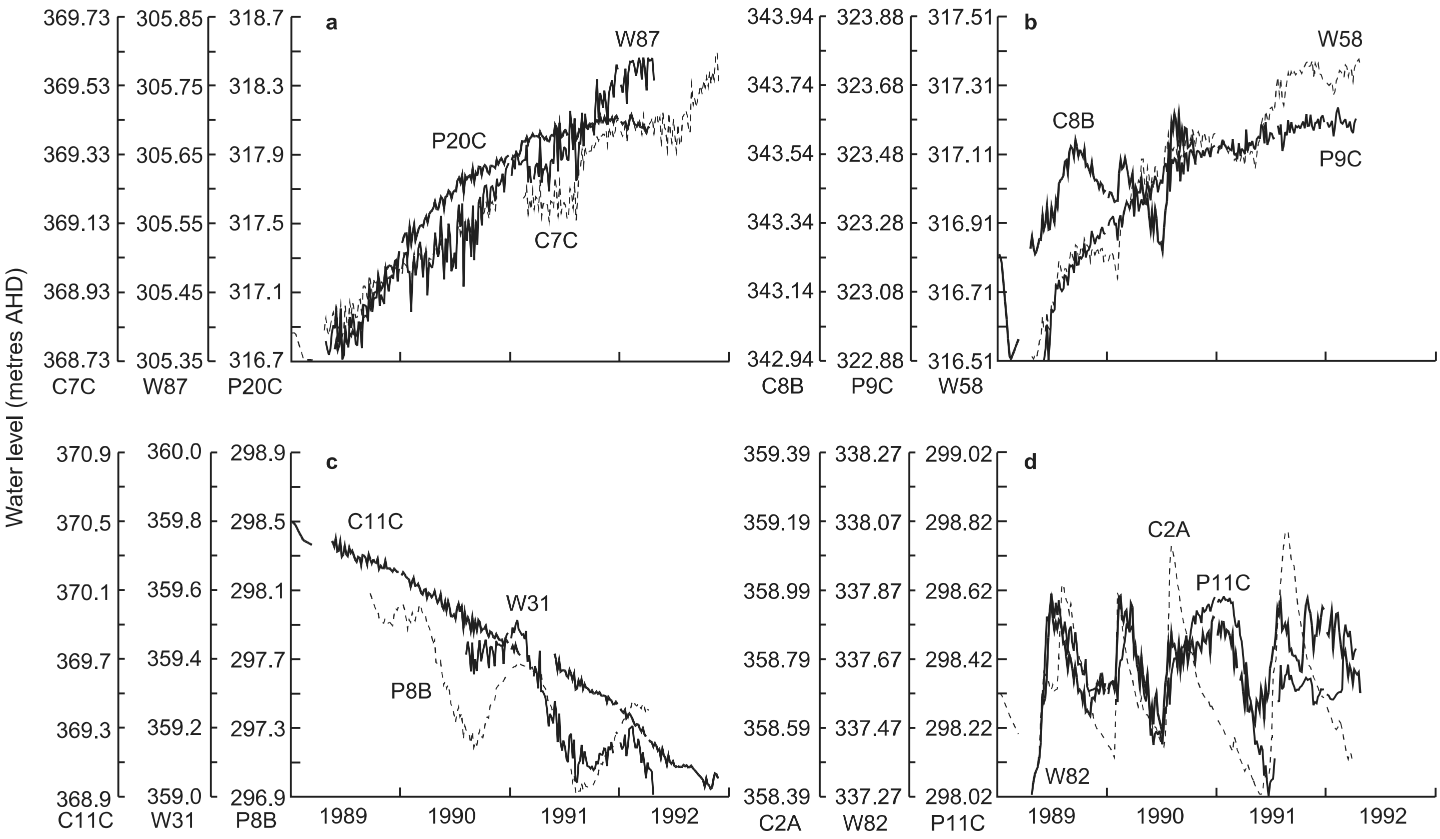


Figure 5a–d Long-term water-level trends for three catchments in the Western Australia wheat belt showing: **a** monotonically rising water levels; **b** continuously rising water levels; **c** continuously declining water levels, and **d** seasonally fluctuating water levels. (After Salama et al. 1993b)

3. Continuously declining water levels (*Figure 5c*), which occur in the watershed areas of catchments where previously non-vegetated areas had been planted by trees.
4. Fluctuating water levels (*Figure 5d*), which occur in the lower reaches of the catchments near streams where seasonal fluctuations in water level are comparable to the rise and decline in stream level. Water levels rise in early winter to reach a peak in early spring in response to the winter rain, and they decline to a minimum during summer.

Groundwater discharge usually takes place when the groundwater levels are converging toward the surface, or when the water levels are above ground level in flat areas and along valleys in the lowlands and at the break of slope in other areas.

Overburden and Related Groundwater-Controlled Soil Formation

Two main geological processes explain the development of the overburden in these catchments (Salama et al. 1993c, 1993d). Both processes are controlled by groundwater. The first is a weathering process, typical of the laterite profile; the profile consists of extensive deeply weathered and altered bedrock zones to a depth of 30 m at the flanks and 50 m in the valleys. The section consists of a mottled zone, a pallid zone, and a weathered bedrock zone, characterized by the widespread presence of a coarse gravelly base, 5–10 m thick, below the valleys.

The second process is deposition, in which the regolith is formed of alluvium, colluvium, sedimentary and lake deposits. The sedimentary deposits are greater than 90 m thick in some areas. The alluvial sequence occupies a very narrow band parallel to the streams and thins toward the flanks. Some areas show a sandplain unit present in the midslopes and underlain by a greyish clay layer. Silicified hardpans are common below the valley flanks and the valley floor. The sedimentary sequence is formed of claystones, sandstones, and conglomerates, ranging in depth from 10–30 m. The lake deposits are formed of salt layers, black clay, and intercalated clastic sediments.

The rock material is rarely affected by one process operating independently, but is usually subject to several interactive processes that act concurrently to produce the end results. For example, whereas weathering might be predominant in the highlands of the catchment, deposition might be predominant in the lowlands.

In both models, layers of calcrete and silicified hardpans usually occur. Two types of calcrete occur in the catchments: pedogenic calcrete, widely distributed in the upper 2 m of the soil; and non-pedogenic calcrete, associated with the drainage system. Silicrete is also widespread and is mainly formed by the rising groundwater. The deposition model is also characterized by the presence of several sedimentary formations and paleochannel sediments mostly preserved along the low drainage lines.

Salinity Distribution

Salinity increases from catchment divides to the valley floors. Chloride concentration in groundwater is much lower in the upper region of the catchments (100–1000 mg/L), where sandplain and pediment slopes are predominant. Groundwater is fresher in these regions, due to the greater amount of recharge and continuous leaching of salts through the transmissive sandy soil. Valleys and paleochannel regions have the greatest salinity (5000–100,000 mg/L), because groundwater moves downstream and salts become concentrated in the channels when water is lost through evaporation. Paleochannels that are not connected to the main stream where salts can be leached act as a salt repository (>50,000 mg/L) (Salama et al. 1993a).

The accumulation of large amounts of salts upstream of geological structures in the valleys indicates a continuous process of groundwater movement from the catchment divide to the valley floors, where water is lost along this route by one or more of the processes explained above. However, nearly all the water that reaches the valleys upstream of the geological structures is lost by diffuse discharge. These relationships are illustrated in *Figure 6*.

Origin of Salts

In Western Australia, detailed studies of atmospheric solute inputs, reported by Hingston and Gailitis (1976) and Farrington et al. (1992), show that chloride concentration in rainfall decreases with increasing distance from the coast. Comparisons between the ionic composition of rainwater and seawater show that terrestrial sources contribute significantly to ions precipitating in rainfall. The sources include airborne salts from saline lakes, fertilizers spread by wheatbelt farmers to grow crops and pastures, and the release of potassium when crop stubble is burnt (Farrington et al. 1992).

Salinity trends, saturation indices, phase diagrams, and mass-balance calculations show that most of the constituents in groundwater can be accounted for by ions in the rain, with some minor additions from the weathering process. Ions are then concentrated in the soil profile by one or more of the previously described four main mechanisms (Salama et al. 1993c).

Mass-balance results (Salama et al. 1993c) suggest that a correlation exists between extent of salinity and the degree of weathering and deposition. Groundwater with the greatest salinity occurs in areas with the greatest deposition, weathering, and cation exchange. In addition, the greatest degree of weathering tends to occur in channels saturated with water. On the basis of these results, two chemical models are proposed (Salama et al. 1993c, 1993d) for the evolution of groundwater in the investigated areas. The first is a weathering model, which is usually noticed in first-order catchments, and the second is a deposition model, present in second- and third-order catchments. The chemical models and the geological processes that control evolution of overburden in the catchments are therefore interrelated.

Weathering Model

Groundwater in first-order catchments in the wheat belt has Na^+ ions in excess of Cl^- ions, and total ions in excess of rain constituents, indicating that weathering is occurring. The weathering produces gibbsite and kaolinite, with the release of Na^+, K^+, Mg^{2+}, Ca^{2+}, HCO_3^-, and H_4SiO_4, in a system that is open to CO_2.

The weathering process is described as follows (Salama et al 1993c):

1. Weathering of the rock minerals through hydrolysis and acid attack occurs to form kaolinite, HCO_3^- and H_4SiO_4, with the release of Na^+, K^+, Ca^{++}, and Mg^{++}.
2. Prior to clearing of forests and reserves, the weathering process was mainly dominated by a system open to CO_2 (supplied by plant roots) and a continuous supply of H^+. This condition stimulated

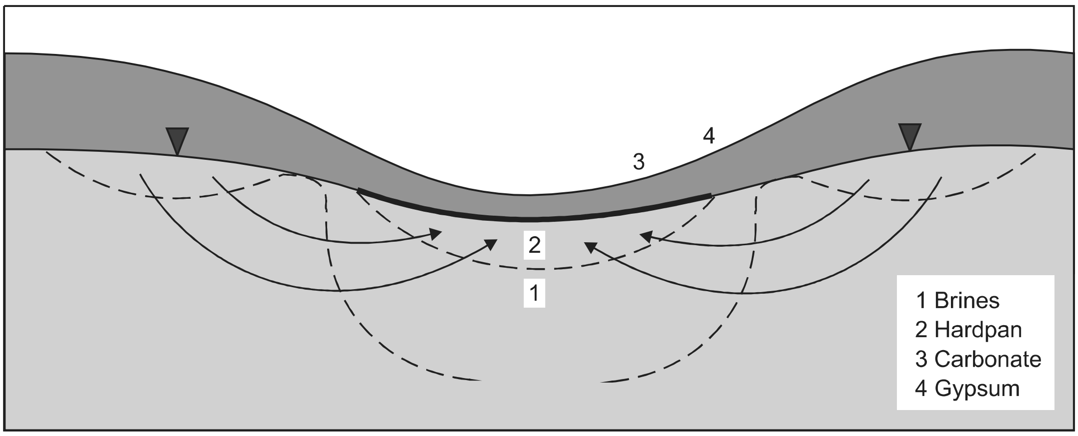


Figure 6 Schematic section showing groundwater flow toward depressions; the accumulation of salt (*1*); and the deposition of hardpans (*2*), carbonate (*3*), and/or gypsum (*4*)

the weathering reaction, causing Na^+ to be released from the feldspar in exchange for H^+ ions. Evidence for this reaction comes from excess Na^+ and high HCO_3^- concentrations in shallow unconfined aquifers and deep aquifers in uncleared forests and reserves.

3. After clearing of trees, the biogenic CO_2 became depleted, causing the H^+ activity and the weathering reaction to decrease. In this new environment, the cation exchange, which was previously taking place as a secondary reaction, now becomes more dominant, causing the water to become depleted in Na^+ relative to chloride.
4. Chemical reactions during the weathering process require substantial amounts of water to convert albite, biotite, hornblende, and K-feldspar to kaolinite, H_4SiO_4, and HCO_3^-. Mass-balance calculations indicate that chemical changes in groundwater could be brought about by weathering reactions with primary minerals albite, biotite, and hornblende in the ratios of 7:3:1. The weathering process is a progressive one. As weathering proceeds, downward erosion of the top layers takes place at about the same rate as the weathering. Using the continuous weathering–erosion process as a basis, one can assume that the uptake of water by the weathering process will be continuous. The newly formed minerals will also be transformed into other mineral assemblages by additional water uptake. For example:

$$\underset{\text{Kaolinite}}{Al_2Si_2O_5(OH)_4}+5H_2O=2H_4SiO_4+\underset{\text{Gibbsite}}{Al_2O_3\cdot 3H_2O} \quad (1)$$

Deposition Model

Chemical deposition or accumulation that applies to the geological deposition scenario described earlier occurs where there is an appreciable loss of the ions brought into the system by rainfall. In this case, groundwater has Cl^- in excess of all other ions. This condition indicates a depositional environment, where the ions contributed to the system by rainfall, together with the weathering products, are used to form new minerals. These minerals are deposited in the form of calcrete and silicified hardpans in the valleys. The calcrete can also become dolomitized. Gypsum is also deposited in depressions and lakes (*Figure 6*).

The deposition model is summarized as follows:

1. Rainwater, having a similar composition to seawater, percolates downward. As the water moves downward, most of the water is removed by the four main mechanisms. Due to these processes, ions accumulate in the unsaturated zone at various levels, depending upon lithological, hydrogeological, and hydrochemical controls (Salama et al. 1993d).
2. Biogenic CO_2 produced by plant roots comes into contact with the setting-up sequence of carbonate equilbria, which involves formation of carbonic acid, bicarbonate, carbonate, and dissolution/precipitation of calcium carbonate:

$$CO_{2(g)}+H_2O=\underset{\text{carbonic acid}}{H_2CO_3} \quad (2)$$

$$H_2CO_3=H^++HCO_3^- \quad (3)$$

$$HCO_3^-=H^++CO_3^{2-} \quad (4)$$

$$Ca+2HCO_3=\underset{\text{calcite}}{CaCO_3}+CO_2+H_2O \quad (5)$$

3. However, precipitation of dolomite is favored by an increase in alkalinity, in the reaction:

$$Mg^{2+}+2HCO_3^-+CaCO_3 =CaMg(CO_3)_2+CO_2+H_2O \quad (6)$$

As Ca is removed, the Mg/Ca ratio increases. At a ratio of 6 or greater (Drever 1982), calcium carbonate (as aragonite) in the sediment is converted to dolomite:

$$2CaCO_3+Mg^{2+}=CaMg(CO_3)_2+Ca^{2+} \quad (7)$$

Here, dolomitization occurs in the absence of excess alkalinity, and the calcium that is released precipitates as gypsum:

$$Ca+SO_4+2H_2O=CaSO_4\cdot 2H_2O \quad (8)$$

The carbonate minerals dolomite, calcite, and aragonite are generally saturated in samples of low salinity. Saturation with respect to gypsum generally increases with increasing salinity.

Saturation with respect to calcite increases under open CO_2 conditions, because the amount of calcite that can be dissolved in water is limited by the availability of CO_2 (Drever 1982). According to Eq. (3), more HCO_3^- is available when CO_2 dissolves in water and this reacts with Ca^{2+} to produce calcite (Eq. (2)). Samples from wells in the sandplain slopes are undersaturated with respect to calcite, because the wells are in recharge areas where mineral concentration is lower in the sandy soil.

4. Mass-balance calculations indicate that magnesium is taken up in the formation of sepiolite, as a reverse weathering process. The mineral is formed from degraded clays and groundwater constituents, according to the equation:

$$Mg^{2+}+2HCO_3+4H_4SIO_4 =\underset{\text{Sepiolite}}{Mg_2Si_3O_6(OH)_2}+2CO_2+6H_2O \quad (9)$$

Carbon dioxide is a product of this reaction, hence releasing CO_2 to compensate for CO_2 consumed in the weathering process.

5. Silicic acid (H_4SiO_4) and other cations produced by the weathering reaction migrate with the groundwater toward the valleys (*Figure 6*). In the valleys, as the groundwater is discharged from the confined aquifers, CO_2 is lost, pH increases, and aluminosili-

cates are precipitated according to the following reaction:

$$K^+ + 2\,Al^{3+} + 1.5\,H_4SIO_4 + 5\,H_2O \rightarrow KAl_2O_3(SIO_2)_{1.5}(H_2O)_5 + 6\,H^+ \quad (10)$$

Increasing aridity, receding water tables, and continuous groundwater discharge lead to the formation of aluminosilicates on the surface. On hardening, this forms the hardpan. Due to the cyclic nature of wet and dry periods, several hardpan layers occur in the valleys (*Figure 6*).

6. Eugster and Hardie (1978) also observe the early removal of potassium in hypersaline environments. This removal is attributed to the interaction of K with clays through adsorption or ion exchange. Clay layers are an extensive sedimentary feature of wheat-belt catchments (Salama and Hawkes 1993), and hence this interaction is a feasible explanation for the observed loss of K from solution.
7. In wet periods and during times of increased rates of recharge, the groundwater usually becomes undersaturated with respect to the carbonate minerals. In this environment, dissolution of previously deposited (usually under drier conditions) carbonate minerals occurs.

Salinity Development in the Landscape

The recharge rate prior to clearing was less than 1 mm yr^{-1} (Otto and Salama 1994; Salama et al. 1993b), and with this low rate of recharge, most of the total soluble salts (TSS) infiltrating with the recharging water is concentrated in the unsaturated zone as the water is distributed by one or more of the four mechanisms. During these periods of low recharge, and where only deep aquifers are present and the groundwater gradients are sluggish, TSS concentration within the aquifers is a continuous process. After clearing, rates of recharge increased 10–20 times (Salama et al. 1993a, 1993b, and 1993e) and new aquifers developed within the unsaturated zone. Consequently, water tables have risen and groundwater gradients and velocities have increased, causing the TSS stored in the unsaturated zone to be mobilized. Thus, more discharge that is saline has taken place at the surface. The TSS distribution pattern suggests that following clearing, TSS mobilization from the divides to the valleys is increasing, resulting in more land being affected by salinity and an increase in salinity in streams.

Case III. Groundwater Controls on Salinity Development in a Paleochannel System (Salt River, Western Australia)

The Salt River, part of the paleodrainage system in Western Australia, developed during Late Cretaceous time and ceased regular flow in inland areas by mid-Miocene time (van de Graaff et al. 1977). A well preserved system of paleochannels with well developed distal fans and paleo-lakes has been reconstructed. Further downstream, the Salt River becomes a series of salt lakes. The total area of the catchment is about 88,000 km^2. Surface-water flow in the Salt River is intermittent except during winter, when the lakes overflow and streamflow is continuous. Although this system forms a distinctly different pattern of salinization compared to that in the smaller-scale catchments of South Australia and Western Australia, the system acts as a salt repository and provides the channels through which salt is transported from the catchments to their final discharge points (i.e., lakes).

Geomorphology and Geology

The valley slopes are formed of sandy plains, rock outcrops, colluvium material, and lateritic duricrust. The sand plains form rounded, gently undulating hillocks associated with alluvial sediments. A lateritic duricrust comprising a hard crust of ferruginous sand and ironstone gravels underlies most of the granitic hills in the northern area of the valley. Salt lakes and playas occupy the downstream part of the valley. The salt lakes are characterized by the presence of thick salt deposits that represent various stages of precipitation, i.e., deposition of sulfates and chlorides.

Development of Salinity

A dramatic decrease in ocean temperature and a concurrent increase in the volume of Antarctic ice occurred in late Miocene and early Pliocene times. These conditions led to a widespread precipitation decrease in inland areas, accompanied by a general increase in aridity associated with continued northward motion of Australia into drier climatic belts (Kemp 1978). The decrease in precipitation and the prevailing aridity caused the formation of sand dunes across the river course in the upstream parts of the Salt River, which led to the formation of a series of inland lakes (Salama et al. 1992; Salama 1994). Alternating wet and dry periods during the Pleistocene Epoch caused the erosion and local deposition of sediments that filled up the river systems, forming paleochannels. Due to the absence of large enough gradients to transport the sediments, the landscape changed gradually into its present form.

The rising Darling Plateau added to the aridity in the wheat-belt zone by reducing the amount of rainfall eastward (Salama 1994); this process led to further drying up of most of the lakes and the formation of evaporites, which were eventually buried by moving sand. This process continued for most of the Tertiary Period. Several wet periods were recorded during this time, which led to the formation of the lateritic duricrust. At later stages, conditions were favorable for the deposition of thick carbonate deposits (Salama 1994). Due to climatic changes from humid to arid, most of the river systems stopped flowing, causing the formation of intermittent pools and internal playas (van de Graaff et al. 1977).

The thick alluvial deposits in the channels acted as a sink for groundwater flow in the catchment. Groundwater flow and its quality varied with the climate change. Salinity profiles along the paleochannels indicate relatively fresher zones within the hypersaline areas. The higher-salinity deposits coincide with the possible dry climatic periods, whereas the fresher zones with vegetative remains coincide with wetter periods.

Case IV. Groundwater Controls on Salinity Development in a Lake System (Lake Deborah, Western Australia)

Lake Deborah represents the final stage of development of salinization in the wheat belt of Western Australia; this stage is characterized by extensive development of salt lakes in closed basins. These lakes form the largest group of salt lakes in the eastern wheat belt. They occupy a relict paleodrainage system, about 30 km in length and 1–9 km in width. The surface area of Lake Deborah is 112 km^2. A halite crust, 10–700 mm thick, underlies the entire surface area of the lake (Salama et al. 1992).

Geomorphology and Geology

Lake Deborah is situated near the center of the Yilgarn Block. This stable Archean craton consists of belts of banded gneiss and layered sedimentary, volcanic, and intrusive rocks that are intruded by voluminous granitoids (Chin and Smith 1983).

The sedimentary sequence of Lake Deborah consists of the following lithostratigraphic units, from top to bottom: (1) hard salt crust; (2) unconsolidated salt layer; (3) greyish-black mud layer (black ooze); (4) clastic sediments; and (5) mudstone layer.

Groundwater flow and chemistry

Lake Deborah acts as a sink for groundwater flow in the catchment (Salama et al. 1992). Groundwater flow into the lake ranges from 480–2880 m^3d^{-1}, based on 48 km of lake perimeter, transmissivity $(T) = 10–30\ m^2d^{-1}$, and hydraulic gradient $= 0.001–0.003$. Of this flow, 250–750 m^3d^{-1} is estimated to be entering through the paleochannel system in the north and the fault system in the south ($T = 30–50\ m^2d^{-1}$; Salama et al. 1992).

Total soluble salts (TSS) of groundwater range from 84,000 mg/L near the northern inlet to 74,000 mg/L in the southern inlet area. Highest TSS near the lake is 100,000 mg/L, compared to 320,000 mg/L for brines within the lake and sediments. Inflow of relatively fresh groundwater into the lake near the northern inlet does not permit the build up of a hard salt crust, as is the case in other areas. This process also occurs at the southern end of the lake, where groundwater flow apparently converges along a fault line. This same phenomenon also causes the salt crust to be thin near the lake boundaries and attain its maximum thickness in the central part.

Groundwater Controls on Salinity Development in the Lake

Analysis of the chemical data (Salama et al. 1992) shows that the brines were formed through continuous evaporation. The relationship between Na and Cl and bromide, which is a conservative element, is constant, as shown in *Figure 7*. During evaporation, Cl and Br concentrations increase steadily until halite saturation is reached. Only a small amount of Br is taken up by the precipitating halite, and hence Br continues to increase at a much greater rate than Cl (Carpenter 1978). Brines for Lake Deborah plot above the seawater trajectory and are therefore interpreted as being formed mainly by continued dissolution of halite. *Figure 8* shows that groundwater samples plot at one end of the line, whereas the lake-brine samples plot at the other end, parallel to the evaporation line; this distribution indicates that the main origin of the brines in the lakes is from groundwater. *Figure 7* shows the same relationship for Na and Br, indicating that no evaporated seawater component exists in the Lake Deborah system. This conclusion is confirmed by the stable-isotope results, shown in *Figure 9*. The samples plot in two distinct fields. The first includes samples from groundwaters south and north of the lake, which are relatively depleted in $\delta^{18}O$ and δD. Brines from the lake plot in the other field. These are highly enriched in δD and $\delta^{18}O$ due to the cyclic pattern of evaporation. Another conclusion from these results is that no migra-

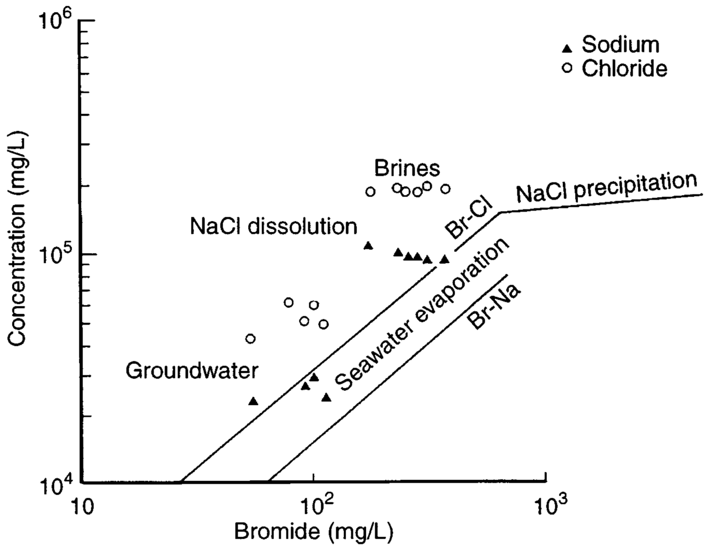


Figure 7 Relation between (*1*) Na and Cl and (*2*) bromide, showing groundwater and lake brines as two distinct groups. (After Salama et al. 1994)

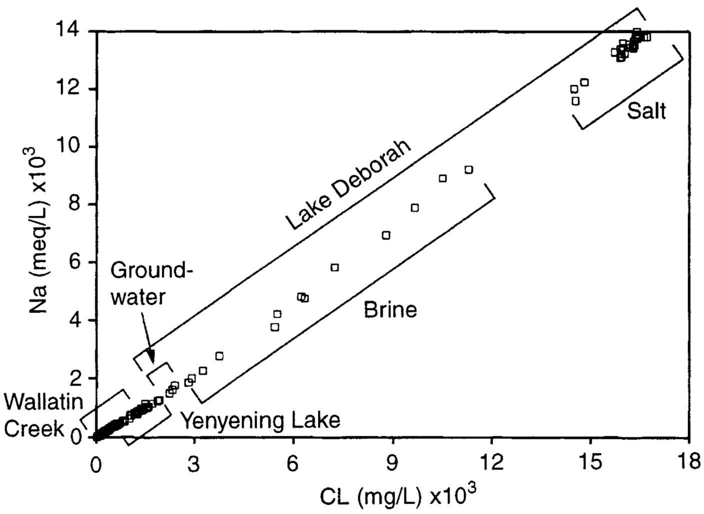


Figure 8 Relation between Na and Cl for Lake Deborah, Yenyening lakes (paleochannel), groundwater samples from near Lake Deborah, Lake Deborah brines, and salt. (After Salama 1994)

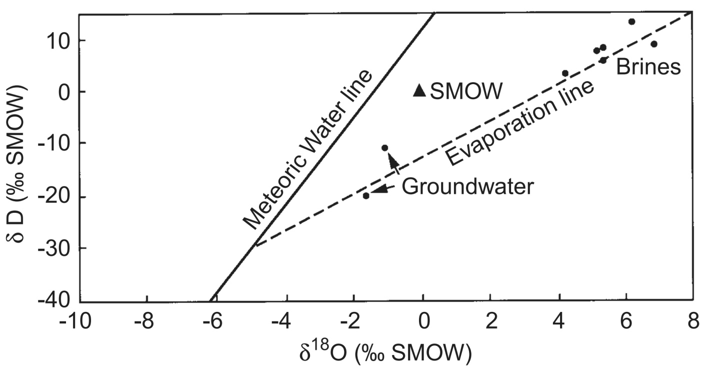


Figure 9 Relations between δD and $\delta^{18}O$ in groundwater samples and in brines from cored wells, showing two distinct fields along the evaporation line. (After Salama et al. 1994)

tion of brines has occurred from the lake toward the groundwater system.

Case V. Groundwater Controls on Salinity and Saline Lakes Development in Regional Aquifer Systems (the Rift Basins of Sudan)

Geomorphology and Geology

The Sudanese rift structures form intra-continental basins that are bordered on all sides by anorogenic terrain (Salama 1997). Locations are shown in *Figure 10*. These basins are the result of a multi-structural system of rifts that have probably been activated several times since the Paleozoic Era. The rapid rates of uplift and subsidence have assisted in rapid accumulation and filling of the basins with unconsolidated sediments, ranging from a few hundred meters to several thousand meters. The rift structures contain sediments of several age groups, origins, and modes of deposition. Some of them are as old as the Paleozoic Era and include sediments of Mesozoic, Tertiary, and Quaternary ages (Bahr El Arab rift and Blue Nile rift). The thickness of the Tertiary sediments exceeds 15 km in the Bahr El Arab rift.

Aquifers

Groundwater in these basins usually occurs in saturated strata of Tertiary and Mesozoic ages. Three types of aquifers are described (Salama 1997):

1. An unconfined aquifer, which underlies most of the basin area, especially in the marginal parts, where recharge takes place from the wadies and along bedrock faults. The unconfined aquifer is not present in the central parts of the basins.
2. A semi-confined to confined aquifer, which occurs mainly in Tertiary sediments.
3. A confined aquifer, mainly in the Mesozoic sediments at the boundaries of the graben.

Evolution of the Groundwater Flow Systems

Based on hydrogeological and hydrochemical data from the rift basins, the general direction of groundwater movement is from the basin boundaries toward the central parts (Salama 1985, 1987). Groundwater troughs occur in each of the rift basins. The water-level depression of the Bahr El Arab basin forms the Sudd trough; 300 m is the lowest level in the trough and the lowest groundwater level in all the rift basins (*Figure 10*). In the White Nile basin, the groundwater movement is from the west to the southeast with a hydraulic gradient of 0.00053; the Nuba trough is at 320 m. The general groundwater movement in the Atabara River basin is from the southwest to the northeast. The Atbara trough is a large trough at the northeastern part of the basin, and a smaller one is present at El Gash. A large groundwater trough exists in the central part of the Blue Nile basin at Gezira and a smaller one at Soba.

Groundwater flow in the rift basins is controlled by four main processes (Verweij 1993): (1) sedimentation in a subsiding sedimentary basin; (2) introduction of heat into a basin; (3) tectonic processes acting on a basin; and (4) infiltration of meteoric water in a subaerial basin. These four processes were at one stage or another working separately or together in the rift basins of Sudan. The rifting phases that formed the Sudanese Rift System caused the formation of deep basins. The continuous subsidence of the basins, together with continuous recharge of meteoric water taking place at the aquifer boundaries (rift-basin boundaries), caused the formation of a deep groundwater trough in each one of these basins. Although no groundwater discharge is presently taking place at these troughs, it is logical to assume that in pluvial periods these troughs would act as groundwater discharge areas.

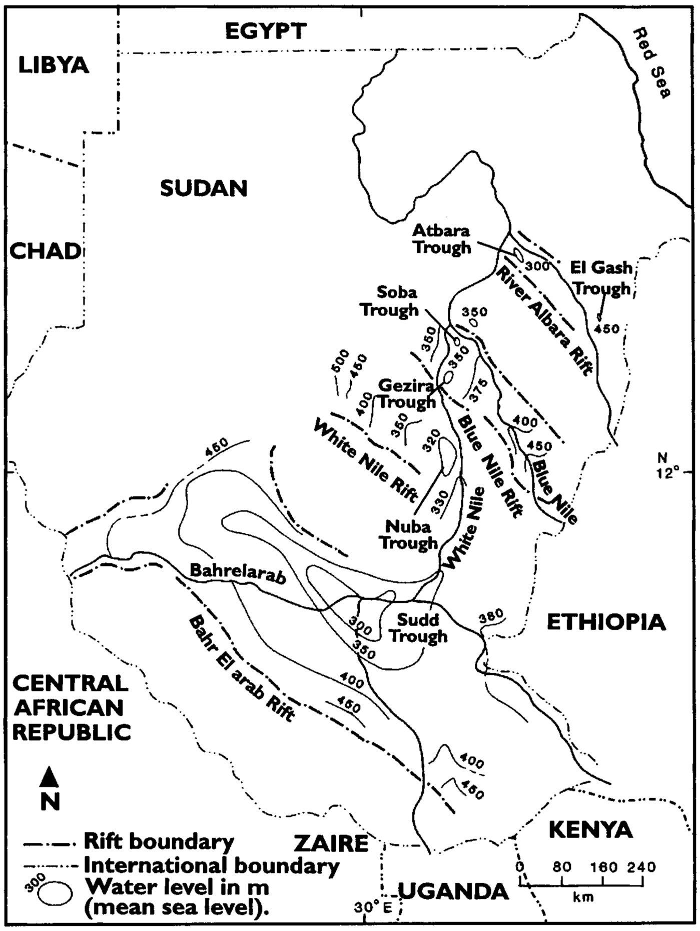


Figure 10 Groundwater levels in the Sudanese Rift basins showing troughs in each rift system. (After Salama et al. 1987)

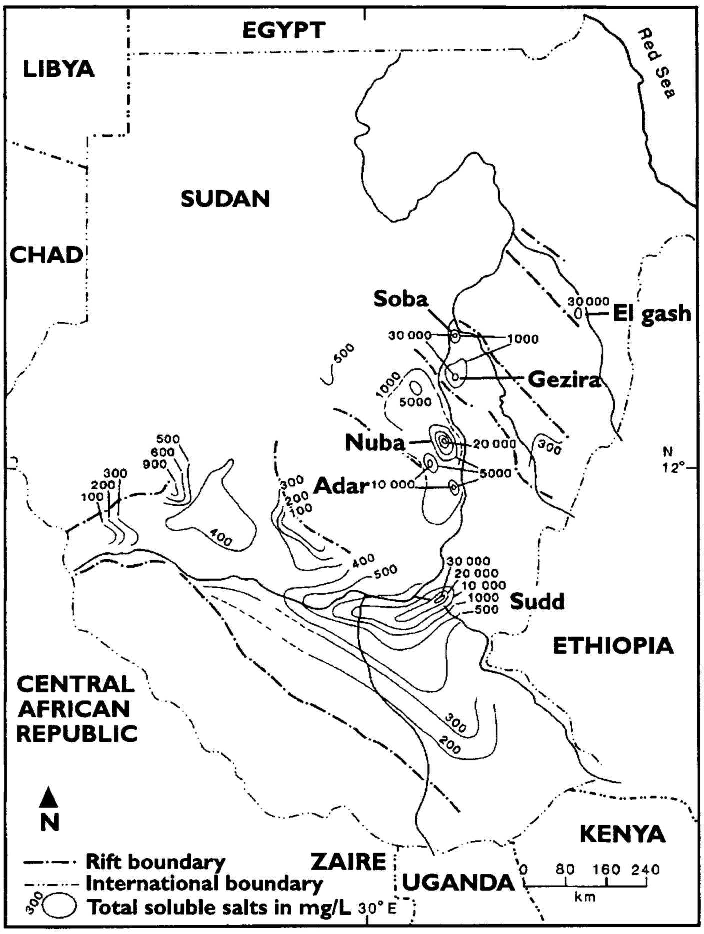


Figure 11 Distribution of total soluble salts and occurrence of saline zones of Sudd. (After Salama et al. 1987)

The continuous subsidence of the basins and the fact that the areas where the troughs currently exist are the deepest parts of the basins, which coincide with the groundwater troughs, indicate that the groundwater occurrence is a replica of the surface-water pattern that existed at the time of formation.

Effect of the Flow Systems on Groundwater Chemistry

Groundwater salinity is generally low at the basin boundaries, as shown in *Figure 11*. Along the western boundary of Bahr El Arab rift, salinity is less than 80 mg/L TDS. Salinity increases gradually as the water moves in the aquifer; values are 500–800 mg/L about 500 km away from the boundary in, e.g., Bahr El Arab basin. This change represents an increase of 1–2 mg/L per km of linear flow in the aquifer. This rate abruptly increases as the saline zones are approached. No decrease in salinity occurs in the vicinity of Bahr El Arab, indicating that recharge or mixing with surface water from Bahr El Arab does not occur. The high minimum salinity of 400 mg/L at the northern boundary of the White Nile rift can be attributed to insufficient surface runoff to leach all the efflorescent salts that were formed in the surface sediments. Alternatively, because of the northerly position of the basin relative to the other basins, the rates of evaporation may have been higher and consequently the initial salinity would be expected to be higher.

In the southern areas of the basin, in a triangular area covering the paleochannel, four highly saline zones are recognized. Salinity in these ridges ranges from 5000–20,000 mg/L. In the southern parts of the White Nile basin, salinity increases from a minimum of 100 mg/L along the eastern boundary of the basin to about 1000 mg/L in the area west of the White Nile. Two highly saline zones are shown that have salinity as great as 15,000 mg/L.

A high-salinity zone extends in the northern part of the Blue Nile basin. This zone is restricted to the lower and upper Tertiary aquifers. Whereas the salinity in the lower Mesozoic aquifer is everywhere less than 500 mg/L, values as great as 30,000 mg/L occur in the Tertiary aquifer in the Soba and Gezira saline zones. These high-saline zones are also restricted to the upper Tertiary sediments.

The saline zone in each rift zone forms a closed ridge, which coincides with the closed groundwater trough (Salama 1985, 1987). The saline body or bodies formed by evaporation coupled with alkaline-earth carbonate precipitation and re-solution of capillary salts. The widespread presence of calcrete, kanker, and other carbonate deposits, overlying and within most of the Tertiary deposits, indicates that conditions were favorable for the deposition of carbonates. It is postulated that the shallow standing waters evaporated, thereby forming salt crusts. In the next flood, fresh water dissolved the most soluble salts (NaCl and Na_2SO_4) and transported these toward the deepest part of the basin, leaving carbonates behind in the form of kanker nodules. This process explains the wide distribution of kanker nodules in the upper Tertiary units and the concentration of sodium, chloride, and sulfates in the saline zones. At the same time, in the deepest part of the graben, which was always a lake, playa, or sebkha, the lake water evaporated and became increasingly saline. During dry periods, the lakes were completely or partly evaporated, thus creating layers of salts that were later dissolved by groundwater to form saline groundwater bodies. The lakes that formed in the central parts of the closed basin fluctuated in size according to the paleoclimatological conditions prevailing at the time.

Case VI. Groundwater Control on Saline Seepage in the Northern Great Plains of North America

The Northern Great Plains Region (NGPR) includes parts of Montana, North Dakota, and South Dakota, USA; and Manitoba, Saskatchewan, and Alberta, Canada. In the NGPR, clayey till provides an excellent soil-moisture reservoir for dryland farming, and the dominant cropping system is alternating grain–fallow farming. Saline seepage is the major soil-salinization problem in the NGPR; this seepage is caused mainly by excess water recharging below the root zone during the fallow period. Other factors contributing water to saline seeps are poor surface drainage, snow accumulation, leaking ponds, artesian water, and crop failure. Salinity in the prairies existed long before the first settlements. Estimates of the amount of land affected by high salinity differ greatly, but probably more than 10% of the arable land is affected by saline seepage (McCracken 1973; McKenzie et al. 1997; Vander Pluym 1978, 1982). A Montana survey shows a 30% increase in dryland saline seeps from 1959 to 1975. An Alberta survey shows a threefold increase in the areas of saline seepage during a 10-year period. Moderate or severe salinization in the Canadian Prairies affects about 1.4 million ha. A salinity-risk index map indicates that 28% of the land has a moderate risk and nearly 10% of the land has a high risk of soil salinization (Eilers et al. 1997).

Climatic Factors

The climate of the NGPR is arid to semi-arid; rainfall is 250–450 mm yr^{-1}. Strong winds remove most of the snow from cultivated upland areas and deposit it in drifts. When the drifts melt in the spring, the soil is readily recharged with meltwater, enhancing seep development. Generally, dryland salinity occurs in areas with more than 350 mm of annual rainfall (Miller et al. 1981). Beneath fallow areas, a water-table rise of 0.05–3.0 m can be expected during wet years. Although the water table declines during the remainder of the year, normally it does not return to the initial level. This indicates a net rise in water-table elevations over the years and continued aggravation of the salinity problem.

Hydrogeomorphic Factors

The landscape is undulating and the shallow bedrock consists of marine and non-marine shales and sandstones (Henry et al. 1985; Miller et al. 1985). The bedrock strata influence the type of material and amount of salt that is incorporated in the overlying glacial drift and associated groundwater (Hendry and Buckland 1990). The unweathered shale and thin bentonite beds create an impermeable layer, inhibiting significant groundwater flow within the units. The low-permeable, weathered shale zone provides a laterally extensive layer that allows groundwater in the overlying glacial till to slowly flow downgradient and to accumulate in low areas or to drain into small streams (Hendry and Buckland 1990; Henry et al. 1985).

The glacial drift has several important effects on the development of soil salinization. The excellent soil-moisture characteristics of the glacial deposits provide effective storage for groundwater, in which excess water accumulates until it reaches the surface as a seep. The poor drainage associated with the rolling topography allows water to pond for long periods of time. Preferred flow paths, such as fractures and joints in the till, allow water to recharge the underlying aquifer. The glacial till contains abundant calcium and sodium salts.

The dominant mechanism of salinization is related to the underlying aquifer and to the movement of groundwater in the catchment (Hendry and Buckland 1990). Hendry et al. (1990) conclude from a study in east-central Alberta that the salinity is most likely related to the seasonally high water tables in the depressional areas. Ponding of surface water and periods of intense rainfall are the causes of the elevated water tables.

Brown et al. (1983) describe seven common types of saline seeps: (1) a geologic outcrop seep occurs when the recharge area is underlain by low-permeable material such as shale, dense till, or clay; (2) a coal-seam seep occurs where recharge areas are underlain by coal or lignite, where the coal seam crops out; (3) a seep develops where the recharge area is underlain by glacial till that overlies more permeable strata of the Tertiary Front Union Formation. Water moves through the glacial till and enters the more permeable strata. In areas where the till or low-permeable strata are truncated by the permeable zone, groundwater ascends to the surface and forms a saline seep; (4) a textural-change seep develops when the texturally coarse-to-medium soil material in the recharge area is underlain by low-permeable material. Water moves to the low-permeable zone and laterally downgradient, where it encounters a lower permeable soil zone, which causes the water table to rise; (5) a slope-change seep occurs in a similar hydrogeomorphic setting as in (1) in areas where the slope gradient decreases, causing sluggish groundwater flow and the water table to rise to the soil surface; (6) a hydrostatic-pressure seep develops in a similar setting as in (1), except that the aquifer becomes confined downslope and the groundwater becomes superhydrostatic, indicating an upward groundwater-flow component to the surface; and (7) a pothole seep occurs in a recharge area that is poorly drained and underlain by low-permeable material. Groundwater moves laterally through a shallow higher permeable aquifer that crops out near the soil surface to form a saline seep.

Groundwater Quality

The overall soluble-salt load contained in each geological formation is reflected in the salinity of the groundwater at the discharge site. A 1978 regional water-quality survey of 2800 sites in Montana (Miller et al. 1981) shows that 14% have TDS concentrations of less than 500 mg/L; 16%, 500–1000 mg/L; 64%, 1000–10,000 mg/L; and 6%, more than 10,000 mg/L. Groundwater in the glaciated Colorado Shale area of Montana (black marine shales) has much higher concentrations of most constituents (Na-Mg-SO_4 are dominant) than in the glaciated Judith–Clagett–Eagle region (Miller et al. 1981). The Colorado and Bearpaw Shale units underlying large areas of the NGPR are loaded with salt, predominantly secondary gypsum, some calcite, and an abundance of other elements. Samples taken from the Tertiary Fort Union units have similar compositions to the Judith–Clagett–Eagle groundwater but are lower in concentration. In general, the occurrence of dryland salinity in areas underlain by these younger units is not extensive at the surface, but areal groundwater salinization is more widespread, suggesting the presence of a deeper, regional flow system. Significant concentrations of trace elements, especially selenium, occurred in many of the groundwater samples. The concentrations are much higher in the Colorado Shale region.

In central Alberta, patterns of groundwater chemistry and soil salinization are similar to flow patterns (Maclean 1974), with increases in TDS, Na, K, SO_4, and Cl downslope in a catchment.

In a saline seep in southern Alberta, precipitation of evaporite minerals is controlled by the solubility of each mineral (Miller et al. 1989). When saline groundwater discharges to the soil surface, less soluble minerals such as calcite and gypsum precipitate within the subsoil, whereas more soluble Na–Mg-type evaporites precipitate on the soil surface. On a regional scale within closed basins, this process can create playas or saline lakes. The shallow groundwater and soil solutions are dominated by Na and SO_4. Gypsum, bassanite and calcite are the only evaporites identified in the subsoil. This evaporite mineral sequence does not follow the Hardie–Eugster model of closed-basin brine evolution. The study suggests that the common-ion effect involving calcite dissolution and gypsum and/or bassanite precipitation is an important mechanism, and that these latter minerals may be a major source of Ca and SO_4 in the groundwater and soil solution.

In east-central Alberta, salt deposits have accumulated from groundwater discharging in a closed topographic depression since the last glaciation (Wallick 1981). The chemical evolution of groundwater is dominated by a shallow and a regional deeper flow system having different types of water–rock interaction. The chemistry of the shallow flow system is oxidizing and aerobic. In the regional bedrock aquifer, conditions are reducing, groundwater is alkaline, and the dominant chemical species are Na, HCO_3, and Cl. Groundwater flow is induced by the postglacial topography, and water from the shallow aquifer mixes with water from the deep bedrock aquifer in the groundwater discharge area, yielding a range of intermediate chemical compositions.

Summary and Conclusions

Groundwater plays a major role in the mobilization, accumulation, and discharge of salts into the landscape. Salinity increases along groundwater flow paths from catchment divides and areas of recharge to the valley floors and areas of groundwater discharge. This pattern is similar in parts of three continents, North America, Australia, and Africa.

In South Australia, saline groundwater discharge activates soil processes of oxidation, reduction, solinization, sulfidization, and sulfuricization, which cause the soils to lose structure and fertility and become susceptible to waterlogging and erosion.

In the hummocky terrains of the NGPR, the localized recharge and discharge scenarios cause salinization to occur mainly in depressions, resulting in the formation of saline soils and seepages. On a regional scale within closed basins, this process can create playas or saline lakes. The shallow groundwater and soil solutions are dominated by Na and SO_4. The chemical evolution of groundwater is dominated by a shallow and a regional deeper flow system having different types of water–rock interaction. Groundwater flow is induced by the postglacial topography; water from the shallow aquifer mixes with water from the deep bedrock aquifer in the groundwater discharge area, yielding a range of intermediate chemical compositions.

In contrast, in Western Australia, recharge prior to clearing was less than 1 mm yr^{-1}, and with this low rate of recharge most of the total soluble salts (TSS) infiltrating with the recharging water are concentrated in the unsaturated zone as the water is distributed by one or more of four mechanisms. During these periods of low recharge, where only deep aquifers are present and the groundwater gradients are sluggish, TSS concentration within the aquifers is a continuous process. After clearing, rates of recharge increased 10–200 times, and new aquifers developed within the former unsaturated zone. Consequently, water tables have risen and groundwater gradients and velocities have increased, thereby causing the TSS stored in the unsaturated zone to be mobilized. Thus, more discharge that is saline has taken place at the surface. The TSS distribution pattern suggests that following clearing, TSS mobilization from the divides to the valleys increases, resulting in more land being affected by salinity and an increase in salinity in streams.

Hydrogeomorphic controls cause salinity to be formed along structural lineaments and barriers, at breaks of slope, in depressions, and along streams. Saline groundwater flows along the beds of the streams and is accumulated in the paleochannel system, which acts as a salt repository. The final discharge points of the saline groundwater are in the lakes, where most of the saline groundwater is concentrated and forms salt crusts, which vary in thickness according to the concentration of the original brine and the prevailing environmental conditions.

In the regional aquifers of the rift basins of Sudan, salinity increases along groundwater flow paths and forms saline zones at the distal ends. The saline zone in each rift forms a closed ridge, which coincides with the closed groundwater trough. The saline body or bodies were formed by evaporation coupled with alkaline-earth carbonate precipitation and dissolution of capillary salts.

In all these areas, from small catchments in the Western Australia wheat belt, through to continental-scale areas in rifts of Sudan, salt has a wide variety of origins. In all the cases described, however, groundwater is the critical geological agent in the development of salinization.

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