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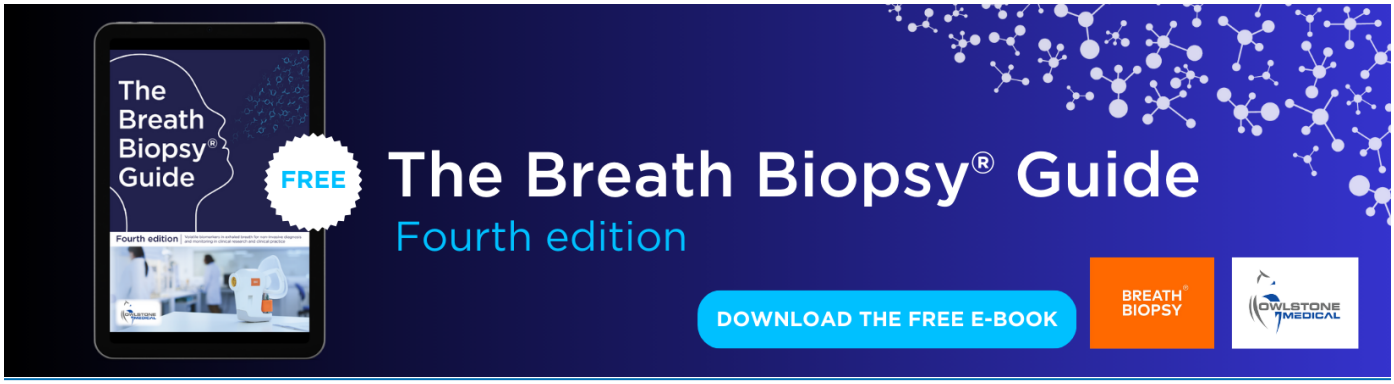

Environmental Research Letters

TOPICAL REVIEW

# Non-renewable groundwater use and groundwater depletion: a review

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## Abstract

Population growth, economic development, and dietary changes have drastically increased the demand for food and water. The resulting expansion of irrigated agriculture into semi-arid areas with limited precipitation and surface water has greatly increased the dependence of irrigated crops on groundwater withdrawal. Also, the increasing number of people living in mega-cities without access to clean surface water or piped drinking water has drastically increased urban groundwater use. The result of these trends has been the steady increase of the use of non-renewable groundwater resources and associated high rates of aquifer depletion around the globe. We present a comprehensive review of the state-of-the-art in research on non-renewable groundwater use and groundwater depletion. We start with a section defining the concepts of non-renewable groundwater, fossil groundwater and groundwater depletion and place these concepts in a hydrogeological perspective. We pay particular attention to the interaction between groundwater withdrawal, recharge and surface water which is critical to understanding sustainable groundwater withdrawal. We provide an overview of methods that have been used to estimate groundwater depletion, followed by an extensive review of global and regional depletion estimates, the adverse impacts of groundwater depletion and the hydroeconomics of groundwater use. We end this review with an outlook for future research based on main research gaps and challenges identified. This review shows that both the estimates of current depletion rates and the future availability of non-renewable groundwater are highly uncertain and that considerable data and research challenges need to be overcome if we hope to reduce this uncertainty in the near future.

## 1. Introduction

Over the last century, the global water cycle has been subject to large changes. The global population has quadrupled, currently exceeding 7 billion, with more than 50% living in urbanized areas (Klein Goldewijk *et al* 2010). The increasing demand for food caused by this population growth (Godfray *et al* 2010) has resulted in a dramatic expansion of irrigated agriculture during the 20th century (Siebert *et al* 2015). Moreover, the rapid urbanization and economic development and associated dietary changes have had a large effect on the water use per capita, both in terms of actual water use as well as the virtual water content of products consumed (Hoekstra and Chapagain 2007). As a result, the abstracted volume of water for human needs has increased from about 500 to ~4000 $km^3\ yr^{-1}$ over the last 100 years (Oki and Kanae 2006, Hanasaki *et al* 2008a, 2008b, Wada *et al* 2014, Hanasaki *et al* 2018).

The expansion of irrigated agriculture into semi-arid areas with limited precipitation and surface water has greatly increased the reliance of irrigated crops on groundwater withdrawal (Siebert *et al* 2010, Wada *et al* 2012a), a trend that has been named the 'silent revolution of intensive groundwater use' (Lamas and Martínez-Santos 2005). This also includes existing irrigated regions that partly rely on surface water for irrigation, e.g. the Central Valley of California, but show increasing trends in groundwater use in response to incidental surface water droughts (Scanlon *et al* 2012a). Moreover, the increasing

number of people living in mega-cities without access to clean surface water or piped drinking water infrastructure (McDonald *et al* 2014) is causing souring rates of urban groundwater withdrawal. The result of these trends has been the steady increase of the use of non-renewable groundwater, i.e. groundwater that is taken out of the aquifers that will likely not be replenished on human time scales (Gleeson *et al* 2012). As a result, the depletion rate of groundwater resources has increased during the last decades (Wada *et al* 2010, Konikow 2011, Wada *et al* 2012a, Döll *et al* 2014, Richey *et al* 2015b, De Graaf *et al* 2017) and is likely to persist in the decades to come (Wada *et al* 2012b).

Although overuse of groundwater was first put on the agenda almost two decades ago (e.g. Postel 1999, Shah *et al* 2000, Lamas and Martínez-Santos 2005, Konikow and Kendy 2005, Giordano 2009), the true extent of non-renewable groundwater use has only become evident to the larger public since 2009, when the first analyses with the GRACE satellite were published that showed persistent negative trends of groundwater storage in heavily irrigated areas in India and Pakistan (e.g. Rodell *et al* 2009), and after the first global assessment of groundwater depletion was published (Wada *et al* 2010). Since then, many additional assessments, either from GRACE (Tiwari *et al* 2009, Longuevergne *et al* 2010, Famiglietti *et al* 2011, Richey *et al* 2015b, Rodell *et al* 2018) or from global hydrological modelling (Bierkens 2015, Wada 2016), were published. These studies produced quite a large variety of results, which, as a consequence, led to extensive scientific discourse about the validity of some of the outcomes and methods used. In many of these publications the terms groundwater depletion, fossil groundwater and non-renewable groundwater are used alternatively without proper definition. Moreover, an additional line of papers appeared that assess the side effects of groundwater depletion, such as increased sea-level rise (Konikow 2011, Wada *et al* 2012b) and regional land subsidence (Sharifi *et al* 2008).

In this review paper, we attempt to provide an overview of the recent research on the use of non-renewable groundwater resources and groundwater depletion and its impacts. Our target audience is the environmental science community at large, rather than the specialised hydrogeologist, groundwater modeller or economist. The review is wider in scope than a recent paper by Wada (2016) that focused mostly on current and future depletion estimates, but less extensive than the book by Margat and Van der Gun (2013) that provides an in-depth overview of groundwater systems around the world. The paper is a classical review where for each theme and topic relevant literature is used to support perceived developments. Apart from interrogating our own archives, we have searched for literature in Scopus and Google Scholar (using search terms related to the topics), and from literature lists of books and papers found previously. We have added a list of themes and topics treated (appendix) with seminal references for each topic. We stress that this not a quantitative review and meta-analysis. Therefore, we do not claim to be complete in our citations. Moreover, the review is quite broad in topics treated, as we aim to provide an informed overview on the important dimensions of non-renewable groundwater use. Because each topic by itself could be subject for a separate review for a specialist audience, it is unavoidable that we had to compromise in terms of depth.

The remaining of this review consist of three parts. The first part (section 2) deviates from the review format as it is meant to define for non-specialists the terms 'non-renewable groundwater', 'fossil groundwater' and 'groundwater depletion' and place these concepts in a hydrogeological perspective. We pay particular attention to the interaction between groundwater withdrawal, recharge and surface water, which is important to understand sustainable groundwater use. The second part consists of the review sections 3–6. It starts with a review of methods that have been used to estimate groundwater depletion (section 3), followed by a comparison of global and regional depletion estimates to obtain a sense of their uncertainty (section 4). This is followed by a review of recent studies assessing the impacts of groundwater depletion (section 5). As groundwater use is by nature an economic activity, we end the review part with providing an overview of theoretical developments in the hydroeconomics of groundwater (section 6). The third part (section 7) provides an outlook for future research based on main research gaps and challenges that are identified in the review.

## 2. Some key definitions and concepts

### 2.1. The dimensions of water use

Before providing definitions and concepts about groundwater per se, it is good to first define the different dimensions of human water use, as it is excessive water use that ultimately results in the depletion of groundwater resources. Water use is a general term that encompasses water demand, water withdrawal and consumptive water use (Döll *et al* 2012, De Graaf *et al* 2014).

*Water demand* is the water that is needed by a specific sector, e.g. domestic, agriculture or industry, to optimize its activities. Domestic demand contains water needed for drinking, cooking, toilet flushing, bathing and watering the garden. Agricultural water demand consists of irrigation water needed for crop growth and livestock feeds, and water directly needed for livestock, i.e. primarily drinking water. Industrial water demand consists of process water for manufacturing and cooling water needed to support production processes. Note that water demand for processing animal products is included in industrial

**Table 1.** Definitions used.

| Term | Definition adopted |
|---|---|
| Fossil groundwater | Groundwater that has recharged before 12000 BP[a]. Relates to the absolute age of groundwater |
| Young groundwater | Groundwater that has recharged after 12000 BP but more than 50 years ago[b]. Relates to the absolute age of groundwater |
| Modern groundwater | Groundwater that has recharged less than 50 years ago[b]. Relates to the absolute age of groundwater |
| Non-renewable groundwater | Groundwater with mean renewal times surpassing human time-scales (>100 years)[c] |
| Extractable groundwater reserve | The volume of groundwater of sufficient quality that can be extracted (technically and economically) from the aquifer at productive rates over a fixed period |
| Physically sustainable groundwater use | Prolonged (multi-annual) withdrawal of groundwater from an aquifer in quantities not exceeding average annual replenishment, resulting in dynamically stable water tables or hydraulic heads |
| Physically non-sustainable groundwater use or groundwater depletion (Called 'overexploitation' by Margat *et al* 2006) | Prolonged (multi-annual) withdrawal of groundwater from an aquifer in quantities exceeding average annual replenishment, leading to a persistent decline in groundwater levels and reduction of groundwater volumes[c] |
| Groundwater mining | Withdrawal of groundwater from an aquifer having predominantly non-renewable groundwater and very small or no replenishment, causing virtually indefinite depletion of aquifer reserves[c] |

[a] Jasechko *et al* (2017).
[b] Gleeson *et al* (2015).
[c] Based on Margat *et al* (2006).

water demand (see Flörke *et al* 2013). Often, water demand is separated into net water demand, i.e. the water demand that is actually needed, and gross water demand including losses. For instance, net water demand for irrigation would be the difference between potential crop transpiration and actual crop transpiration without irrigation. Gross water demand is then equal to net water demand plus the percolation and evaporation losses that occur during water transport and water application during irrigation.

*Water withdrawal* is the amount of water that is taken from surface water or groundwater to satisfy gross water demand. In case water availability is sufficient, water withdrawal is equal to gross water demand. In case water availability is not sufficient, water withdrawal is equal to water availability. In this case one has a *water gap* equal to gross water demand minus water availability. Note that in certain analyses (e.g. Yates *et al* 2005), recycling of water is taken into account (mostly in the industrial sector). In this case water withdrawal can also be smaller than gross demand as part of the demand is met from the recycled water. In addition, in some water scarce regions, deficit irrigation often takes place and water is supplied less than optimally leading to reduced gross water demand.

*Consumptive water use* is the amount of withdrawn water that is lost due to evaporation (including crop transpiration). The remaining part, i.e. water withdrawal minus consumptive water use, is called *return flow* to groundwater and/or surface water and is available for water use elsewhere but subject to water quality concerns or wastewater treatment in many cases. Note that even if water availability is sufficient, consumptive water use is not the same as net water demand because net water demand only includes the useful part of the evaporated water and not the evaporation losses due to e.g. transport.

### 2.2. Fossil groundwater, non-renewable groundwater and groundwater depletion

In quite a few papers, including those written by the authors of this review, the terms fossil groundwater and non-renewable groundwater and depletion are used interchangeably, while they are not strictly the same. We will provide definitions here and illustrate these with some schematics. Table 1 summarizes the various definitions introduced. Similar definitions were given previously by Margat *et al* (2006, table 1 therein), Margat and Van der Gun (2013, section 4.6 therein), Gleeson *et al* (2015) and Jasechko *et al* (2017).

#### *2.2.1. Fossil groundwater*

The term 'fossil' in fossil groundwater pertains to groundwater of a certain age that surpasses human history. With 'age' we mean the time that has passed since a drop of groundwater has recharged, i.e. since a drop of water from rainfall or from river bed leakage has entered the groundwater body. The age at which groundwater is called 'fossil' is not strictly defined. We follow Jasechko *et al* (2017) who define all groundwater that pre-dates the beginning of the Holocene (approximately 12 000 years B.P.) as fossil. Gleeson *et al* (2015) further subdivide the non-fossil

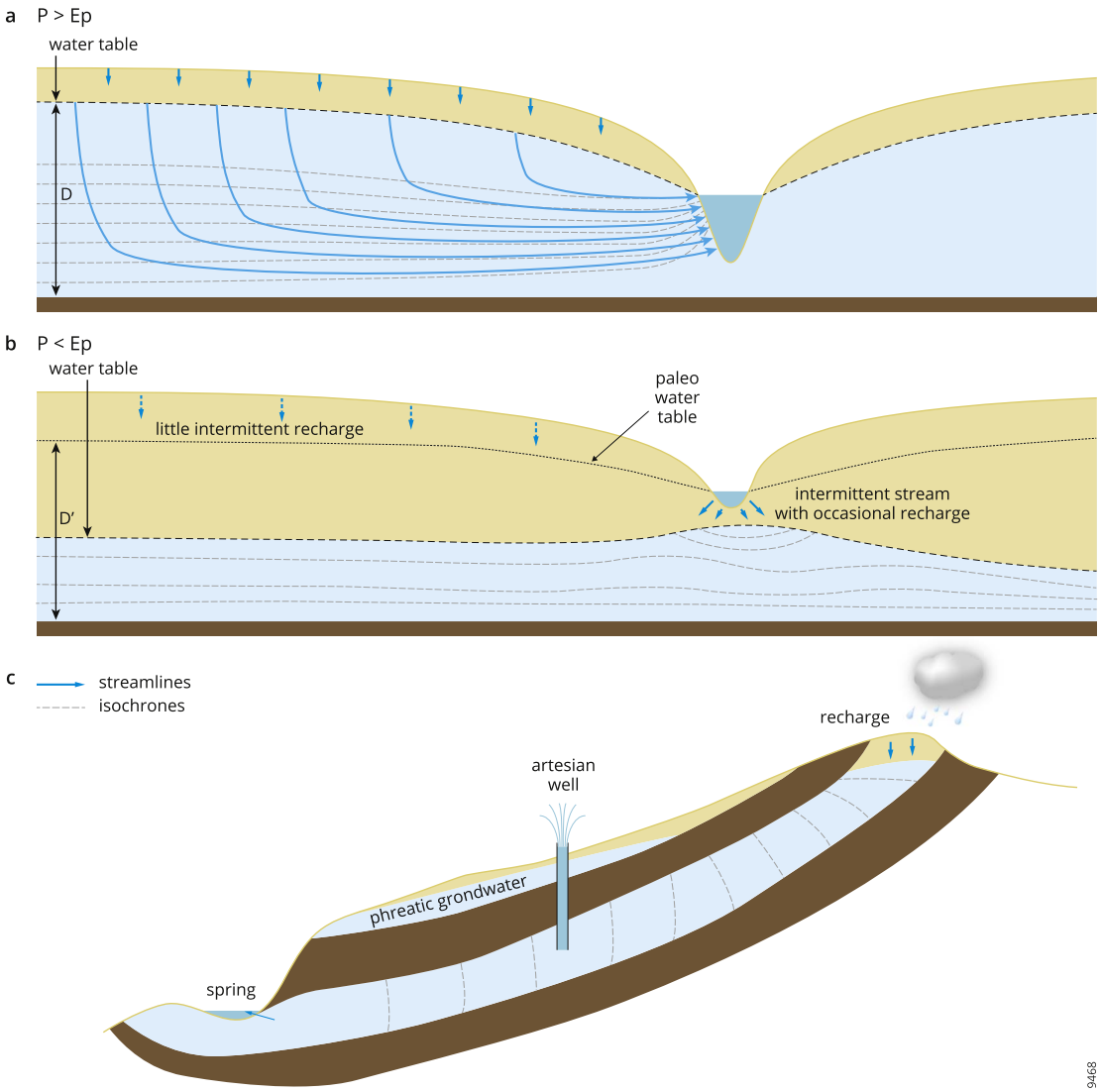


**Figure 1.** Examples of groundwater systems with different age distributions; (a): in a phreatic aquifer in a humid region with a draining groundwater system; (b): in a similar groundwater system under semi-arid to arid conditions; (c): in a confined aquifer system. Groundwater in case (a) is renewable and in cases (b) and (c) non-renewable.

groundwater (younger than 12000 BP) into modern groundwater (younger than 50 years) and young groundwater (between fossil and modern). We will also follow this suggestion.

The occurrence of fossil groundwater greatly varies between groundwater systems. Figure 1 provides three schematic examples of groundwater systems showing lines of equal groundwater age, or isochrones. Figure 1(a) is an archetypical phreatic groundwater system with a free groundwater surface is present, called the water table. In case of a humid to semi-humid climate, i.e. long-term precipitation is larger than long-term potential evaporation, a hydrologically active groundwater body develops that drains to the surface water system. If the aquifer is homogeneous and isotropic, i.e. hydraulic conductivity is the same everywhere and independent of direction, then the isochrones are almost horizontal, except close to the surface water. In this case, a simple equation describing groundwater age as a function of depth can be derived (Ernst 1973, Broers 2004):

$$t = \frac{nD}{R} \ln\left(\frac{D}{D-z}\right). \quad (1)$$

With $t$ groundwater age (years), $n$ drainable porosity (−), $R$ recharge (m yr$^{-1}$), $D$ aquifer thickness (m) and $z$ depth below water table (m). This shows that theoretically, very old and thus fossil groundwater is present even in this most elementary groundwater system, but only close to the bottom of the aquifer. Obviously, in reality aquifers are heterogeneous and anisotropic while recharge rates are generally not homogenously distributed across the surface. So, although groundwater age generally increases with depth, groundwater ages for a given depth may vary considerably.

Figure 1(b) shows the same aquifer, but now in a semi-arid to arid climate, where long-term average precipitation is smaller than long-term potential evaporation. In this case, water tables are deeper and the

aquifer is only occasionally recharged following short periods of heavy rainfall and from leaky streams. In such systems, groundwater reserves are relicts from times that recharge was much larger. Examples are the Nubian aquifer in Africa (Margat and Van der Gun 2013) that was last recharged during the African Humid Period between 15 and 5 ka BP (Claussen *et al* 2017). In this system, the groundwater age reflects the system with larger recharge of the past. Equation (1) still applies, but with $D$ and $z$ with respect to the palaeo-water table and the time since the termination of the humid period added. In case of the Nubian Aquifer, there is also evidence of alternating dry and wet periods over the last 200 ka (Castañeda *et al* 2009) and probably before as dating with krypton isotopes revealed groundwater ages of order $10^6$ years from samples taken at 1200 m depth from the Nubian aquifer system (Sturchio *et al* 2004).

Figure 1(c) provides an example of a confined aquifer system with a phreatic system on top. Confined aquifers can be recharged from humid areas where deeper permeable geological layers, e.g. sand stones, limestone, or unconsolidated gravelly or sandy deposits, are in contact with the surface. If these recharge areas are at topographically higher levels than the larger part of the aquifer, the groundwater is under pressure and will rise above the surface when punctured by a well (artesian well). The location of a groundwater sample from such an aquifer may well be several tenths to hundreds of km away from the recharge location, resulting in very old groundwater ages. An example is the Great Artesian Basin in Australia, where groundwater ages are found to vary between 20 Ka and 1.5 Ma based on chlorine 36 dating (Torgersen *et al* 1991). Note that the aquitard, i.e. the less permeable layer on top of the confined aquifer, may also be partly leaky, which would in the case of over pressure bring the fossil groundwater much closer to the surface where it will mix with the younger groundwater in the phreatic aquifer above. In fact, Jasechko *et al* (2017) showed that fossil groundwater, with an age of over 12 ka, is the dominant groundwater type below 200 m depth, for most aquifers around the world.

#### 2.2.2. Non-renewable groundwater

In order to define non-renewable groundwater, the concept of mean renewal time is often used (Margat *et al* 2006). Mean renewal time is defined as the volume of the groundwater in stock (stored in the aquifer) divided by the mean recharge rate. Thus, large mean renewal rates occur in case groundwater stocks are very large or recharge rates are very small or both. For instance, the Nubian Sandstone Aquifer system in Northern Africa with an estimated volume of more than half a million $km^3$ and recharge rates of a few millimetre per year has a mean renewal time of over 75 000 years (Margat and Van der Gun 2013). Figure 1(a) shows an example of renewable groundwater, and figures 1(b), (c) of non-renewable groundwater. Margat *et al* (2006) provides more insightful examples of the occurrence of non-renewable groundwater in different geographical settings. In Margat and Van der Gun (2013) (based on Richts *et al* 2011), the geographical distribution of the large aquifer systems with non-renewable groundwater are given as well as a table with their estimated volumes and mean renewal times. Note that no absolute age is attached to non-renewable groundwater. Instead, a time-scale is attached to its mean renewability. Figure 1(c) shows that non-renewable groundwater and fossil groundwater are related but not the same. The term 'non-renewable' pertains to the aquifer system as a whole, while age is a location property. A confined aquifer system such as figure 1(c) may consist of non-renewable groundwater, but only part of its groundwater is fossil.

The distinction between renewable from non-renewable groundwater is partly arbitrary. Margat *et al* (2006) suggest a mean renewal time of 1000 years. Here, we will use the following definition (see table 1): non-renewable groundwater is groundwater with mean renewal times surpassing human time-scales. What is meant with 'human time-scale' is again not well-defined. We choose the time-scale of 100 years, which is the limit of human lifetime to date. This shorter time has been chosen as it relates better to the time horizon of groundwater users. If this groundwater is taken out of storage it will not be replenished for the current generations which makes it in effect non-renewable.

The difference between renewable and non-renewable groundwater also impacts its use. If groundwater is renewable, i.e. volumes are small and/or replenishment rates are high, groundwater withdrawal rates would, at least initially, be determined by replenishment rates. When largely non-renewable, groundwater exploitation is determined by the *extractable reserve*, i.e. the volume of groundwater of sufficient quality that can be extracted (technically and economically) from the aquifer at productive rates over a prolonged period.

#### 2.2.3. Groundwater depletion

Following Margat *et al* (2006), we define *groundwater depletion* as the prolonged (multi-annual) withdrawal of groundwater from an aquifer in quantities exceeding average annual replenishment, leading to a persistent decline in groundwater levels and reduction of groundwater volumes. In case there is virtually no groundwater recharge occurring, non-renewable groundwater withdrawal leads to almost indefinite depletion and is also called *groundwater mining* (table 1). We note that this definition is a steady-state view on groundwater depletion. In reality, increased withdrawal in the dry growing season often causes large depletion rates, followed by partial recovery of the water table from subsequent recharge in the wet season. Also, multi-year variability in dry and wet years

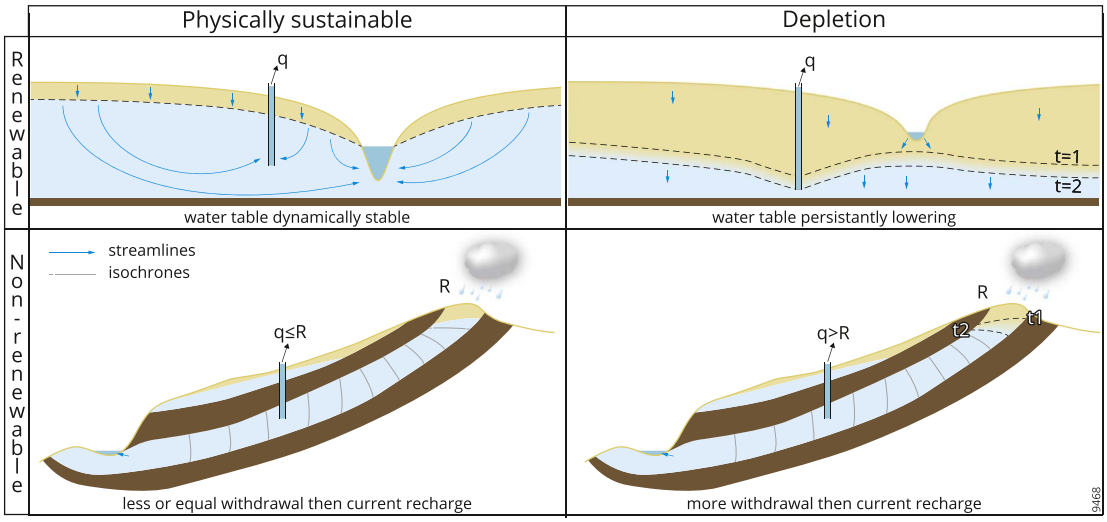


**Figure 2.** Dimensions of groundwater withdrawal: renewable (upper row) and non-renewable (lower row); physically sustainable groundwater withdrawal (left column) and groundwater depletion (right column).

may cause periods of depletion to alternate with periods of recovery (Scanlon *et al* 2012a). In this case a fluctuation pattern of groundwater storage change is superimposed on a long-term declining trend. Our definition of groundwater depletion pertains to that long-term trend.

By definition, groundwater depletion can occur in aquifers with renewable and non-renewable groundwater resources (figure 2). Figure 2 (upper left) shows and example of groundwater withdrawal in a humid system with renewable groundwater. Some groundwater is first taken out of storage, creating a depression cone, but after a while all of the pumped groundwater consists of capturing recharge that would otherwise have ended up discharging to the stream (see also section 2.3). Groundwater withdrawal leads to (dynamically) stable groundwater levels and is therefore physically sustainable[5]. If the well is not too deep, most of the groundwater that is pumped will be young groundwater. Figure 2 (upper right) shows an example of a phreatic aquifer in a semi-arid to semi-humid climate where pumping exceeds the long-year average recharge from streams and rainfall events. Here, groundwater can in principle be still renewable (if pumping stops, recovery could occur in a time scale of years to decades), but over-exploitation leads to depletion. In terms of age, the groundwater pumped will be young or modern. As the water table declines under persistent over-exploitation, the water pumped will become progressively older and be called fossil at some point. Figure 2 (lower left) shows the situation that non-renewable (and fossil) groundwater is pumped from a confined aquifer, but at rates that are smaller than the aquifer's recharge. If withdrawal from the aquifer exceeds recharge, groundwater is depleted (Figure 2 lower right).

[5] We use the term *physically* sustainable, because withdrawal could at the same time lead to reduced streamflow affecting aquatic ecosystems (see section 2.3).

### 2.3. Groundwater-surface water interaction and the sources of pumped groundwater

For deep confined aquifers with little to no surface water interaction (figures 1(b), 2 (lower rows)), the degree of groundwater depletion is only dependent on the balance between recharge and withdrawal. However, groundwater depletion of phreatic aquifers under humid to semi-humid conditions also depends on groundwater-surface water interaction. This fact is often neglected in studies (including the authors' first: Wada *et al* 2010), that wrongfully assume that the maximum allowable withdrawal rate that is physically sustainable is equal to groundwater recharge. This is called the 'water budget myth' by Bredehoeft (1997) and we refer to e.g. Sophocleous (1997), Alley and Leake (2004) and Zhou (2009) for critical comments on and reviews of the subject. In this section, we describe groundwater-surface water interaction and its consequences for physically sustainable groundwater withdrawal. In a classic paper, Theis (1940) explained the source of groundwater pumped and our explanation is based on his work and later explanations by e.g. Alley *et al* (1999), Bredehoeft (2002) and Konikow and Leake (2014).

Under pristine conditions we can divide the interactions between streamflow and groundwater into two main categories: gaining streams where groundwater discharge is contributing significantly to streamflow (figure 3(a)) and loosing streams where groundwater is replenished by infiltrating water from the stream which adds to the recharge from precipitation (figures 3(b) and (c)). The category 'loosing streams' can be sub-divided into two subcategories; connected losing streams (figure 3(b)), when the groundwater level and river bed are connected and infiltration rate

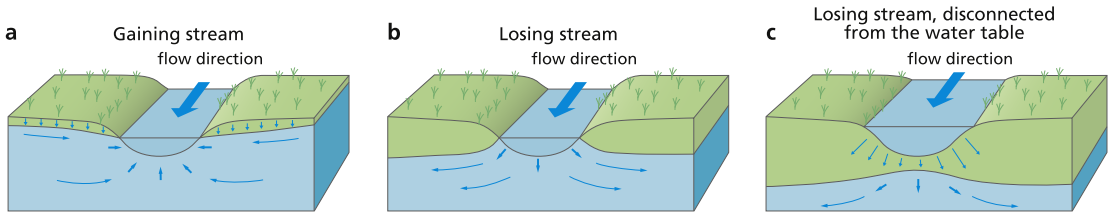


**Figure 3.** Groundwater-streamflow interaction: (a) gaining stream; (b) losing stream; (c) losing stream disconnected from the water table; modified from Winter *et al* (1998); credit to the Unites States Geological Survey.

varies with difference in groundwater and river levels, and disconnected losing streams (figure 3(c)), where the groundwater level is not connected to the river bed, and infiltration loss from the stream mainly depends on river level and quickly becomes constant at greater groundwater depths. This can be readily seen by calculating sensitivities of the infiltration flux to surface water and groundwater levels using Darcy's law in figures 3(b) and (c). Denoting $I$ the infiltration flux, $s$ the surface water level and $h$ the groundwater level, it follows that in case of a connected stream (figure 3(b)) the infiltration flux is equally sensitive to $s$ and $h$: $\frac{\partial I}{\partial s} = \frac{\partial I}{\partial h} \equiv$ constant, while in case of a disconnected stream we have (figure 3(c)): $\frac{\partial I}{\partial s} \propto \frac{1}{h}$ and $\frac{\partial I}{\partial h} \propto \frac{s}{h^2}$.

Draining, connected losing and disconnected losing streams may occur simultaneously at different locations within one catchment, or at the same location at different times of the year. Gaining streams are dominant in (semi-)humid climates and wet periods, and in the lowest parts of a catchment where groundwater that infiltrated higher up in the catchment convergences to. Losing streams are typical for climates and periods when potential evaporation is large enough to evaporate most of the rainfall and for locations higher up in the catchment with very permeable soils. The start of groundwater withdrawal may reverse the flow of rivers from a gaining to a losing stream.

What happens to groundwater-streamflow dynamics under groundwater withdrawal is schematically depicted in figure 4. We start in figure 4(a) with the pristine situation of a gaining stream. Figures 4(b)–(d) show what happens under increasing withdrawal rates. These figures may pertain to one location where withdrawal rates are increasing over time, or multiple locations with similar hydrogeological setup with different withdrawal rates. Note that the following description is done from the perspective of constant meteorological forcing and withdrawal rates, which in reality of course will fluctuate over the years and between years. It thus provides a long-term average perspective on groundwater-surface water interaction.

Figure 4(b) shows a withdrawal regime where the withdrawal rate ($q_1$) is limited. Just after withdrawal starts, the main source of water to a well comes out of groundwater storage of the aquifer. When time passes, the water table starts to develop a gradient towards the pumping well and part of the groundwater recharge, that otherwise would have supplied water to the stream, now contributes to the pumped water (figure 4(b.1)). Consequently, groundwater discharge (*GD* in figure 4) to the stream is reduced. Also, evapotranspiration (*E* in figure 4) from groundwater dependent vegetation will decrease due to falling groundwater levels. The deeper the water table, the larger the contribution of reduced groundwater discharge and reduced evaporation (together termed *increased capture*[6]) to pumped water becomes, until a new equilibrium is reached where all pumped water comes out of capture (see figure 4(b.2)) and groundwater levels stabilize. Thus, withdrawal rate $q_1$ is physically sustainable. Under limited withdrawal rates, groundwater discharge will remain positive and the stream remains a gaining one, albeit with lower streamflow.

If the withdrawal rate is higher than in figure 4(b) ($q_2$ figure 4(c)), groundwater level decline is larger, evaporation is further reduced, and groundwater discharge may shift to surface water infiltration making the stream a losing stream (figure 4(c.1)). However, as long as the groundwater level and river are connected, the contribution of recharge from the infiltrating river water increases with falling groundwater levels. Due to this negative feedback, a new equilibrium will be reached even under $q_2$, where all pumped water comes from captured recharge, streamflow infiltration and decreased evapotranspiration (figure 4(c.2)). Consequently, withdrawal rate $q_2$ is still physically sustainable.

If the withdrawal rate is even higher ($q_3$ figure 4(d)) and groundwater levels drop below the river bed, groundwater gets disconnected from the stream. This situation can be seen as a critical threshold, as the stream recharge rate remains almost constant when groundwater levels drop further (or more precisely, approaches a constant flux with declining groundwater levels). The negative feedback that limits water table decline in case of a connected stream is no longer present, which induces an acceleration of water table decline and successive reduction of evaporation.

[6] Called *increased capture* because the well captures groundwater recharge from rainfall and loosing streams that would otherwise have contributed to streamflow and evaporation.

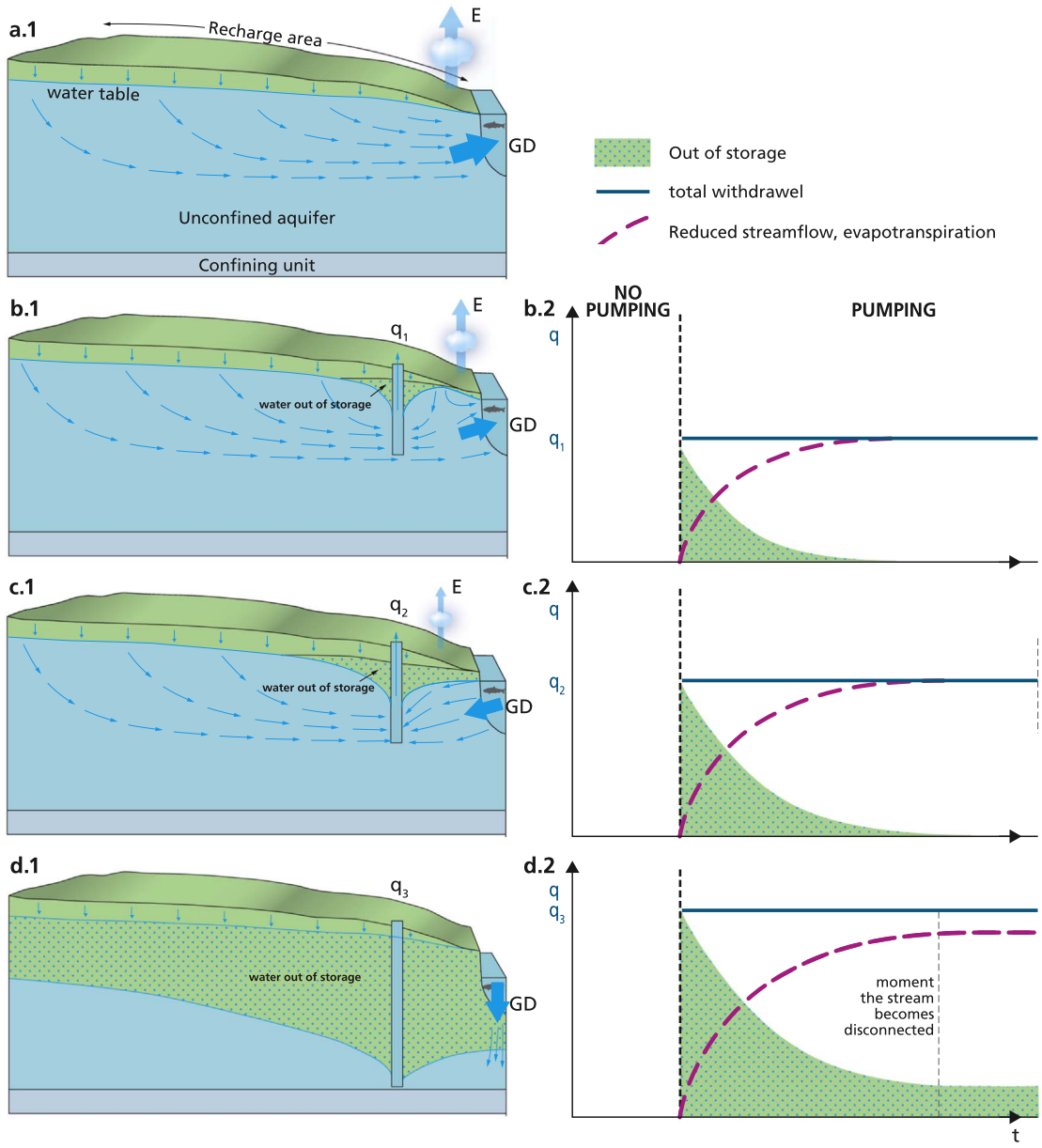


**Figure 4.** Schematic representation of groundwater-surface water interaction under groundwater withdrawal; (a) gaining stream, natural conditions; (b) gaining stream, limited withdrawal. Limited groundwater withdrawal leads to reduced groundwater discharge and evaporation. Groundwater is taken out of storage at first, but eventually a new equilibrium is reached (b2) where all pumped groundwater comes from reduced groundwater discharge and evapotranspiration; (c) connected losing stream, higher withdrawal rates. With higher withdrawal rates, more water is taken out of storage, but again a new equilibrium is reached (c.2). However, here groundwater withdrawal leads to reversal of groundwater discharge and a losing stream; (d) disconnected losing, even more intense groundwater withdrawal rates. Intensive groundwater withdrawal leads to a disconnection of the stream and the water table. As a result, infiltration from the stream reaches a maximum value independent of groundwater level (d.2). If withdrawal rates are higher than infiltration from the stream and diffuse recharge over the depression cone, water is persistently taken out of storage leading to a continuous lowering of the water table at a faster rate; modified from Alley *et al* (1999) and Konikow and Leake (2014); credit to the United States Geological Survey.

Moreover, if the withdrawal rate becomes higher than the maximum stream infiltration rate and the recharge over the depression cone (as often the case in areas with intensive groundwater dependent irrigation), the excess rate of withdrawal comes out of the aquifer storage. As a result, groundwater levels will continue to decline and groundwater storage is persistently being depleted (figure 4(d.2)).

To conclude, similar to evaporation where in a drying soil two stages of evaporation can be distinguished: an energy limited stage where evaporation remains more or less constant and a water limited stage where evaporation deceases in time with diminishing soil moisture a (Philip 1957, Richie 1972), we can also define two stages of groundwater withdrawal in phreatic aquifers: stage 1 (figures 4(b) and (c)) where groundwater is connected with the surface water system, groundwater depletion is limited, and groundwater withdrawal mostly diminishes streamflow and evaporation—one could further distinguish stage 1a and stage

1b by distinguishing gaining and loosing streams-, and stage 2 (figure 4(d)) where further withdrawal in excess of recharge and (constant) stream water infiltration mainly leads to groundwater depletion and does not further impact streamflow.

The actual response time for the water levels to reach a new equilibrium in stage 1 withdrawal depends on hydrogeological properties and dimensions of the aquifer and boundary conditions (river levels, distance well from the stream) and can sometimes be multiple decades (Konikow and Leake 2014). Given that groundwater withdrawal for irrigation often occurs in semi-arid areas with little to no precipitation surplus during the growing season (Wada *et al* 2010), stage 2 is prominent for these regions and progressive groundwater depletion is the rule. We also note that in many regions of the world groundwater is pumped from deeper (semi-)confined aquifers (De Graaf *et al* 2017, figure 1(c)). Under confined conditions groundwater-streamflow interaction only occurs for the larger rivers that are deep enough to penetrate the confining layer. Also, because storage coefficients of (semi-)confined aquifers are much smaller than the drainable porosity of phreatic aquifers, hydraulic head declines in (semi-) confined aquifers are often much larger than water table declines in phreatic aquifers.

## 3. Methods to assess groundwater depletion

The assessment of groundwater depletion is non-trivial because of the limited data available for measuring the state of underground water (Taylor *et al* 2013). When considering methods to assess depletion we can distinguish between volume-based methods, water balance methods and indirect geodetic estimates. In this section, a number of assessment methods falling into these categories are described and examples from the literature given. Note that we do not provide an extensive literature review of depletion estimates here, as this will be given in section 4 which provides an overview of recent estimates of regional and global groundwater depletion. The methods described here are summarized in table 2 in terms of resolution, spatial extent, accuracy and data availability. Figure 5 provides examples of each of the methods described.

### 3.1. Volume-based methods

Volume based methods directly estimate the change in stored groundwater volume over time. These methods are generally the most accurate, because they implicitly take account of increased capture that may occur during groundwater withdrawal (see section 2.3). Konikow (2011) provides global estimates of groundwater depletion based on a large number of regional estimates using volume-based methods. Examples of volume-based methods are:

#### *3.1.1. Volume change estimates based on hydraulic head data*

The most direct observation of the change in groundwater storage is measuring with a piezometer. Changes in groundwater levels (phreatic aquifers) or hydraulic head (confined aquifers) are directly measured with monitoring wells. Local storage change is assessed by multiplying with drainable porosity (phreatic aquifers) or the groundwater storage coefficient (confined aquifers). The accuracy of water table elevation or hydraulic head measurements is high, so that the accuracy of storage change mainly depends on the accuracy of the estimates of drainable porosities or storage coefficients. Hydraulic head data essentially provide point estimates. To obtain regional estimates, one needs to spatially interpolate or average storage changes at points (e.g. Scanlon *et al* 2012b, MacDonald *et al* 2016). The accuracy then depends on the number of observations in relation to the spatial variability of storage change (i.e. the spatial sampling error). Figure 5(a) provides an example of interpolated storage declines from Scanlon *et al* (2012b).

#### *3.1.2. Volume change estimates based on remote sensing with Gravity Recovery and Climate Experiment (GRACE)*

The GRACE mission consisted of twin circumpolar satellites that measured their mutual distance and altitude continuously. From these observations anomalies in the earth gravity field were derived that are mostly attributable to water storage changes on land. The satellites were active between 2002 and 2017 yielding almost 15 years of valuable data on global terrestrial water storage. A follow-up mission called GRACE-FO was launched on 22 May 2018 extending the existing time series. When GRACE gravity anomalies are used for estimating groundwater storage change, the anomalies need to be adjusted first for changes in atmospheric moisture, glacier mass loss, soil moisture and surface water storage. To obtain these terms, estimates from atmospheric circulation models, ice models, hydrological models or additional satellite observations (e.g. soil moisture, surface water levels) are used to isolate groundwater storage change. The accuracy of GRACE total water storage estimates is quite high (order 10–30 mm water equivalent) (Scanlon *et al* 2016), but the accuracy of estimated groundwater storage changes is generally much lower because of the correction with the other storage terms. The temporal resolution of GRACE is monthly (to achieve complete global coverage) and the spatial resolution about 300 $\times$ 300 km. This means that only large regions, aquifer systems or river basins can be monitored adequately. The first published estimates of regional groundwater depletion came in around 2009 (Rodell *et al* 2009, Tiwari *et al* 2009). Only recently, a complete global assessment of terrestrial water storage trends, including groundwater depletion, was published (Rodell *et al* 2018, see figure 5(b)).

**Table 2.** An overview of methods that can be used to estimate groundwater depletion; accuracy estimates from: explanation symbols: − − very low/small; − low/small; o neutral/average; + high/large; + + very high/large.

| | Method | Resolution | Extent | Accuracy | Data-availability |
|---|---|---|---|---|---|
| Volume based methods | Hydraulic head data | + + | + | + + | − − |
| | | Point scale, daily | Regional extent possible in case of averaging wells | 1–10 mm depending on accuracy porosity and sampling error | Often under embargo Poor global coverage |
| | GRACE | − − | ++ | + | ++ |
| | | 300 × 300 km monthly | Globally available | 10–30[a] mm depending basin size and removal other water storage signals | Freely available |
| | Groundwater models | + | ++ | o | + |
| | | Varies widely from 10 × 10 km globally to 25 × 25 m regionally | Globally available | Dependent on hydrogeological, head and discharge data varying between 0.5 and 10 m[b] | Availability depends on model group |
| Water balance methods | (Global) Hydrological models | + | ++ | − − | + |
| | | Varies widely from 10 × 10 km globally to 25 × 25 m regionally | Globally available | 15%–25%[c] But does not account for increased capture of streamflow and evaporation | Availability depends on model group |
| | Remote sensing of fluxes (+models) | + + | + | − − | o |
| | | 1 × 1 km | Regional studies | 20%[d] But does not account for increased capture of streamflow and evaporation | Availability unknown |
| Indirect Geodetic methods | GPS measurements (subsidence or uplift) | + + | + | o | o |
| | | Points, daily | Regional studies | Unknown | Availability unknown |

[a] Scanlon *et al* (2016).
[b] De Graaf *et al* (2017)
.[c] Wada *et al* (2010).
[d] Cheema *et al* (2014).

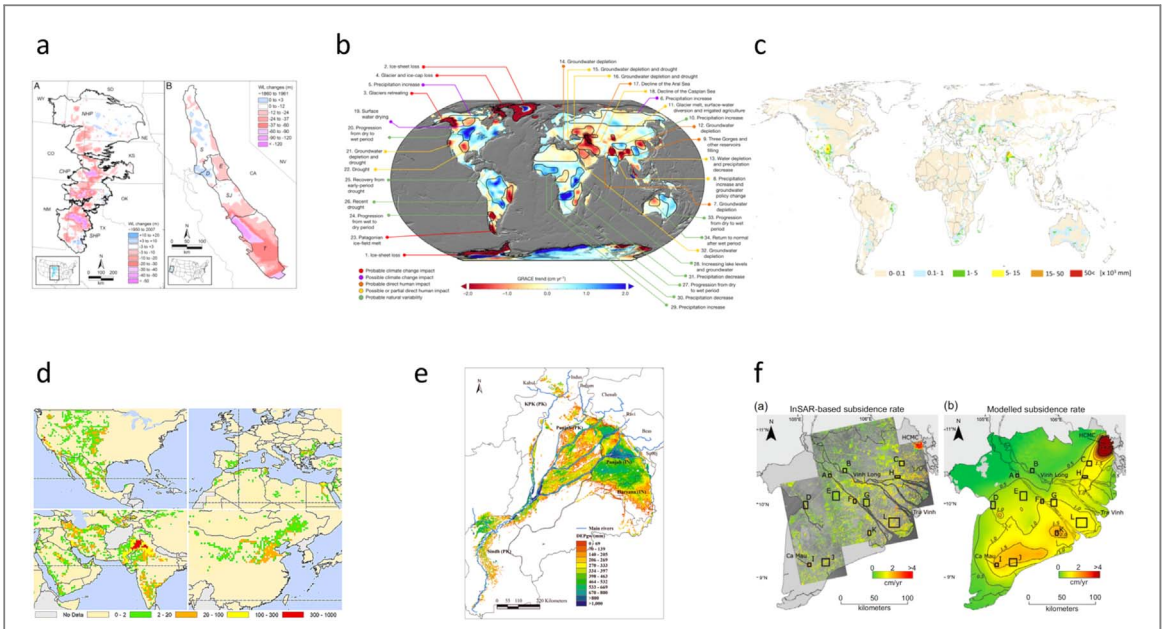


**Figure 5.** Examples of estimates of groundwater depletion using different methods as described in table 2; (a)–(c) volume based estimates; (d), (e) water balance estimates; (f) indirect estimates using geodetic data; All figures reproduced with permission; (a) Groundwater depletion in the High Plains aquifer and Central Valley aquifer estimated from interpolated hydraulic head (Scanlon *et al* 2012b; Panel A from McGuire 2009; credit to USGS); (b) global estimate of terrestrial water storage trends 2002–2015 using GRACE. Part of the declines are the result of groundwater depletion (Rodell *et al* 2018); (c) Global groundwater depletion estimated from a global groundwater flow model (De Graaf *et al* 2017); (d) Global groundwater depletion estimated with the global hydrological model PCR-GLOBWB (Wada *et al* 2010); (e) Groundwater depletion estimates for the Indus Basin using remotely sensed precipitation and evaporation the hydrological model SWAT (Cheema *et al* 2014); (f) Land subsidence rates in the Mekong delta (Vietnam) estimated using a groundwater and geomechanical model (right panel) compared with InSar data (left panel) (Minderhoud *et al* 2017).

#### *3.1.3. Volume change estimates based on (global) groundwater models*

Groundwater flow models are routinely made to support regional (e.g. Faunt 2009, Oude Essink *et al* 2010) or national groundwater management (Henriksen *et al* 2003, De Lange *et al* 2014), while recently the first transient global-scale groundwater model was introduced (De Graaf *et al* 2017). Groundwater flow models are able to calculate groundwater depletion rates in space and time based on simulated groundwater level and head declines multiplied with respectively drainable porosity and storage coefficients. Although volume-based, groundwater depletion estimates with groundwater flow models are only useful in case of a proper modelling of groundwater-surface water interaction and the enhancing effect of shallow groundwater levels on capillary rise and evaporation, (in order to correctly account for increased capture; section 2.3). Moreover, the estimated volume of depletion is strongly dependent on additional datasets, such as groundwater recharge and groundwater withdrawal. These are often calculated from hydrology and water resources models (e.g. De Graaf *et al* 2014, Sutanudjaja *et al* 2018) which may be subject to considerable uncertainty. The spatial resolution of groundwater models varies largely and so is the accuracy that can be achieved (see table 2). Figure 5(c) provides an example of a global map of total depleted volume since 1960 (mm water equivalent) as obtained from De Graaf *et al* (2017).

### 3.2. Water balance methods

Water balance methods compute the groundwater withdrawal rates and compare these with recharge rates. In case withdrawal rates are larger than recharge, groundwater depletion is assumed to occur. The problem with water balance methods is that they typically include diffuse recharge from soils but ignore recharge from water bodies and streams (i.e. focused recharge). Also, they do not account for increased capture (see section 2.3) of streamflow and evaporation (Theis 1940, Konikow and Leake 2014). As recharge is difficult to measure and groundwater withdrawal rates are largely unknown, large-scale hydrological models often use water balance methods to assess groundwater depletion. Here, two methods are discussed, one using hydrological models and one using a combination of remote sensing information and models.

#### *3.2.1. Depletion estimates based on (global) hydrological models*

Many hydrological models have a loss term denoting recharge to (deeper) groundwater or they model groundwater as a simple reservoir with a storage-outflow relationship (e.g. Van Beek *et al* 2011). If groundwater withdrawal data are available, groundwater depletion is estimated by subtracting withdrawal rates from recharge rates and setting depletion equal to the difference if negative. In this case, return flow to groundwater from irrigation is added to recharge rates

$$\text{depletion} = \min(\text{recharge}(\text{natural} + \text{return flow}) - \text{withdrawal}, 0). \quad (2)$$

The first global assessment of groundwater depletion using this method was published by Wada *et al* (2010) (figure 5(d)) using the global hydrological model PCR-GLOBWB (Van Beek *et al* 2011, Wada *et al* 2011). Assessments were made at 30 arc-minutes resolution, which required downscaling of country-based withdrawal data from IGRAC (www.igrac.net). This was done by calculating surface water availability and total gross water demand using PCR-GLOBWB and attributing country-based groundwater withdrawal proportional to the difference between gross water demand and surface water availability. As pointed out by Konikow (2011), this approach does not properly account for increased capture, leading to an over-estimation of groundwater depletion. Wada *et al* (2012a) acknowledged this by using a correction factor based on the volume-based estimates collected by Konikow (2011). In later work (De Graaf *et al* 2014, Döll *et al* 2014) groundwater depletion was calculated with water use schemes that explicitly included surface-groundwater interaction and return flows leading to lower estimates of global groundwater depletion than reported by Wada *et al* (2010).

*3.2.2. Based on remote sensing of fluxes (+ models)*

From remote sensing, it is possible to assess both precipitation (e.g. Huffman *et al* 2007) and evaporation (e.g. Bastiaanssen *et al* 2012). In case evaporation is larger than precipitation, one may assume that the area is irrigated. In case surface water supply is calculated with a hydrological model, the remaining irrigation water should come from groundwater withdrawal. After calculation of groundwater recharge, again with a hydrological model, it is then possible to assess groundwater depletion. This approach, using the SWAT model (Arnold *et al* 1998) and remotely sensed precipitation and evaporation, was followed by Cheema *et al* (2014) to assess groundwater withdrawal and depletion for the Indus basin at very high (1 $km^2$) resolution (figure 5(e)). At this resolution, it is however questionable if lateral groundwater flow and groundwater-surface water interactions can be ignored. Thus, similar to a full model-based approach just described above, depletion rates may be over-estimated.

### 3.3. Indirect geodetic estimates

One of the effects of groundwater depletion is land subsidence (Galloway and Burbey 2011). It is caused by decreased pore pressure as a result of head decline and subsequent consolidation of (mostly) soft sediments, such as can be found in deltas, valley fills, floodplains and former lake beds. Examples of land subsidence as a result of groundwater withdrawal can be found in e.g. Mexico City (Ortega-Guerrero *et al* 1999), California (Amos *et al* 2014), China (Chai *et al* 2004), the Mekong Delta (Minderhoud *et al* 2017), Iran (Motagh *et al* 2008) and Jakarta (Hay-Man Ng *et al* 2012). Estimates of land subsidence rates use geodetic methods. Traditionally this has been surveying, but lately remote sensing methods such as GPS, airborne and space borne radar and lidar are frequently applied. Also, a number of studies mentioned use a combination of geomechanical modelling (e.g. Ortega-Guerrero *et al* 1999, Minderhoud *et al* 2017) and geodetic data to explain the main drivers of land subsidence. Mostly these have been combined in a forward approach, whereby a geomechanical model is driven with known groundwater head declines or withdrawal rates and validated with the geodetic data. A few papers (e.g. Zhang and Burbey 2016) use a geomechanical model together with a withdrawal data and geodetic observations to estimate hydraulic and geomechanical subsoil properties. However, there are no studies known to us that attempt to reconstruct groundwater depletion itself from inverse modelling of land subsidence. This would however be an interesting new and possibly accurate way to estimate groundwater depletion rates at high relative resolution and large spatial extent. Figure 4(f) shows an example of modelled and observed head declines from Minderhoud *et al* (2017).

## 4. Estimates of groundwater withdrawal, groundwater depletion and groundwater storage

This section provides a review of estimates of groundwater withdrawal, groundwater depletion and groundwater storage that have appeared in the literature. Here, we limit our review to global to large-scale studies (e.g. continents, large aquifers, countries) in order to provide a worldwide view, in contrast to e.g. Custodio (2002) who presents a collection of smaller case studies in USA, Mexico, China, Spain, and some Middle East and North African countries.

### 4.1. Global estimates of groundwater withdrawal

Thus far, at the global scale, several studies have attempted to estimate groundwater withdrawal based on two different approaches: country reporting/inventories and modelling.

*4.1.1. Country reporting and inventories*

Despite limited information, a few previous studies gathered reported groundwater withdrawal information for major groundwater users (e.g. Zektser and Everett 2004, Shah 2005). These data-based estimates are primarily based on available country statistics, ranging from 600 to 800 $km^3\ yr^{-1}$ globally (around the year 2000). In addition to these studies, the International Groundwater Resources Assessment Centre (IGRAC) provides a comprehensive database of groundwater related information that includes

country groundwater withdrawal rates worldwide. This database (GGIS; https://un-igrac.org/global-groundwater-information-system-ggis) includes a wide variety of groundwater data for various countries, transboundary aquifer systems, and small islands. These data-based estimates generally rely on country level government reports based on local and regional measurements of groundwater withdrawals, where available. However, they tend to contain many missing data in regions such as Asia, Africa, and South America, where a considerable part of groundwater withdrawals may remain unreported. Although substantial efforts have been made to produce such estimates worldwide, the difficulty remains to update the groundwater withdrawals that are affected by both socio-economic development, and climate variability and change over time. For example, the IGRAC GGIS database reports groundwater withdrawal for India to be 190 $km^3$ (around year 2000), while Foster and Loucks (2006) suggest 240 $km^3$ for later years.

It should also be noted that groundwater withdrawal varies considerably within a countries, while within-country withdrawal estimates are barely available. For the USA, the United States Geological Survey (USGS) provides county level water use data including groundwater withdrawals every 5 years, based on country wide census estimates (https://water.usgs.gov/watuse/). For some other countries, reported province level groundwater withdrawal estimates are also available: Mexico (CONAGUA; https://gob.mx/conagua/), India (Central Ground Water Board; http://cgwb.gov.in/) and China (Ministry of Ecology and Environment; http://english.mee.gov.cn/). These reported withdrawal data have been used in earlier work to validate regional groundwater withdrawals simulated with global hydrological models (see e.g. figure 6 of Wada and Heinrich 2013). Obtaining reliable estimates of groundwater withdrawal worldwide remains, however, difficult, which is even more the case for consumptive use, i.e. withdrawal minus return flows, as it is hardly measurable at large scales (Ruddell 2018).

#### *4.1.2. Hydrological modelling*

Given the many missing data, non-reported groundwater withdrawals, and the difficulty updating withdrawal statistics annually, hydrological model simulation has been widely used as an alternative to estimate groundwater withdrawals and depletion worldwide. Döll (2009) and Döll *et al* (2012) explicitly used information on 'fractional groundwater to total water withdrawals per country' from the IGRAC GGIS database and estimated global groundwater withdrawal to be 1100–1500 $km^3\ yr^{-1}$ (around year 2005). Later, Döll *et al* (2014) used more detailed information on irrigated areas equipped with surface water or groundwater irrigation based on more than 15 000 national and subnational administrative units (Siebert *et al* 2010) and estimated global groundwater withdrawal to be about 700 $km^3\ yr^{-1}$ (around year 2005). This approach is, however, heavily affected by simulated water withdrawal volume per country by using the fractional groundwater to total water withdrawals. Wada *et al* (2010) combined available global-scale information of country groundwater withdrawal rates obtained from the IGRAC GGIS and downscaled them with gridded water demands and availability simulated by a global hydrological model, thus constraining estimated global groundwater withdrawal (800 $km^3\ yr^{-1}$) with reported estimates. De Graaf *et al* (2014) used a dynamic attribution approach based on model-based estimated of the local availability of surface water and groundwater to distribute water demand over surface water and groundwater withdrawal. This resulted in estimated groundwater withdrawals of 909 $km^3\ yr^{-1}$ (year 2000) and 1067 $km^3\ yr^{-1}$ (year 2010). Hanasaki *et al* (2018) used a similar approach as Döll *et al* (2012), i.e. estimating the fractional contribution of surface and groundwater withdrawal, but for each water use sector separately. They estimated global groundwater withdrawal to be $789 \pm 30\ km^3\ yr^{-1}$ (year 2000) based on detailed water withdrawal simulations per water source, including streamflow ($1786 \pm 23$), aqueduct water transfer ($199 \pm 10$), local reservoirs ($106 \pm 5$), and seawater desalination ($1.8 \pm 0$), respectively.

Compared to reported data, these model-based estimates have the clear advantage of having global coverage, but they often neglect physical, technological and socio-economic limitations in water withdrawals that exist in various countries (Wada *et al* 2014). Potential errors can be substantial, given the considerable variation among the estimates (Wada 2016). The range of estimated global groundwater withdrawals in recent studies is, however, constrained between 700 and 900 $km^3\ yr^{-1}$ (around the year 2000). Table 3 provides an overview of some recent estimates of global groundwater withdrawal.

### 4.2. Global estimates of groundwater depletion

As explained in section 3, global estimates of groundwater depletion can be divided into water balance methods and volume-based methods.

#### *4.2.1. Water balance methods*

Global estimates of groundwater depletion are mostly limited to water balance methods (Wada 2016). Postel (1999) provides one of the earliest estimates of 200 $km^3\ yr^{-1}$ (contemporary), which is based on extrapolated country statistics. Most water balance methods however rely on global hydrological models (Bierkens 2015). Earlier model studies (e.g. Vörösmarty *et al* 2005, Rost *et al* 2008, Hanasaki *et al* 2010, Wisser *et al* 2010, Pokhrel *et al* 2012a, 2012b, Yoshikawa *et al* 2014) used the difference between human water demand of agriculture, industry and households and surface water availability as a

**Table 3.** Global estimates of groundwater withdrawal and depletion ($km^3\ yr^{-1}$) (updated from Wada 2016).

| References | Groundwater withdrawal/ depletion $km^3\ yr^{-1}$ | Year | Notes | Sources |
|---|---|---|---|---|
| Postel (1999) | NA/200 | Contemporary | Data-based estimate | Literature and reports<br>Country statistics |
| Shah (2005) | 800–1000/NA | Contemporary | Data-based estimate | Literature and reports<br>Country statistics |
| Zektser and Everett (2004) | 600–700/NA | Contemporary | Data-based estimate | Literature and reports<br>Country statistics |
| Wada *et al* (2010) | 312 (±37)/126 (±32) | 1960 | Model based water balance method | IGRAC GGIS database |
| | 734 (±82)/283 (±40) | 2000 | | PCR-GLOBWB (0.5°) |
| Konikow (2011) | NA/145 (±39) | 2000–2008 | Model and GRACE based volume based method with extrapolation for other than USA, north India, North China Plain, Saudi Arabia, Nubian and Sahara | *In situ* groundwater level measurements, GRACE satellite observation, calibrated groundwater model, extrapolation (15.4%; depletion to abstraction ratio of USA) |
| Wada *et al* (2012) | 312 (±37)/64 (±16) | 1960 | Model based water balance method with correction against reported regional depletion estimates | IGRAC-GGIS |
| | 734 (±82)/204 (±30) | 2000 | | PCR-GLOBWB (0.5°) |
| | 1248 (±118)/295 (±47) | 2050 | | |
| Pokhrel *et al* (2012a, 2012b) | NA/455 (±42) | 2000 | Model based water balance method | MATSIRO (1.0°) |
| Döll *et al* (2014) | NA/113 | 2000–2009 | Model based water balance method | WaterGAP (0.5°), *In situ* groundwater level measurements, GRACE satellite observation |
| Van Dijk *et al* (2014) | NA/92 | 2003–2012 | GRACE based volume based method with data assimilation (original depletion equals 168 $km^3\ yr^{-1}$) | Data assimilation with GRACE satellite observation |
| Wada and Bierkens (2014) | 372/90 | 1960 | Model based water balance method | IGRAC-GGIS |
| | 952/304 | 2010 | | PCR-GLOBWB (0.5°) |
| | 1621(±128)/597(±85) | 2099 | | |
| Yoshikawa *et al* (2014) | −/510 | 2000 | Model based water balance method | H08 (1.0°) |
| | −/1150 | 2050 | | |
| Pokhrel *et al* (2015) | 570/330 | 2000 | Model based water balance method | MATSIRO (1.0°) |
| De Graaf *et al* (2017) | 460/NA | 1960 | Model based volume based method | PCR-GLOBWB (0.083 33°) coupled to a global two-layer MODFLOW model |
| | 980/NA | 2010 | | |
| | NA/137 | 1960–2010 | | |
| Hanasaki *et al* (2018) | 789 (±37)/182 (±26) | 2000 | Model based water balance method | H08 (0.5°) |

proxy for 'non-renewable or non-local water resources', resulting in estimates ranging from 400 to 1700 $km^3$ $yr^{-1}$ (around year 2000). Here, it is important to note that 'non-renewable or non-local water resources' are not further specified and may consist of a combination of non-renewable groundwater withdrawal, water diversion and desalinated water. As a result, these estimates vary substantially between studies.

The first model-based global estimate of groundwater depletion was produced by Wada *et al* (2010). They reported that global groundwater depletion increased from 126 (±32) to 283 (±40) $km^3$ $yr^{-1}$ from 1960 to 2000. The analysis was limited to sub-humid to arid climate zones to avoid overestimation arising from increased capture of discharge and enhanced recharge due to groundwater withdrawal (Bredehoeft 2002). Later, Wada *et al* (2012a, 2012b) applied a correction factor to constrain the original groundwater depletion estimate (Wada *et al* 2010) by regionally reported numbers, producing a 30% lower estimate (204 ± 30 $km^3$ $yr^{-1}$). Döll *et al* (2014) combined hydrological modelling with information from well observations and GRACE satellites, and simulated focused groundwater recharge from surface water bodies in dry regions, while Wada *et al* (2010, 2012a, 2012b) included only diffuse recharge. Moreover, Döll *et al* (2014) applied a deficit irrigation scheme (assuming an irrigation gift of 70% of the maximum water requirements) to constrain their agricultural water withdrawal and estimated global groundwater depletion to be 113 $km^3$ $yr^{-1}$ (average of 2000–2009). Pokhrel *et al* (2015) used an integrated hydrologic model, which explicitly simulates groundwater dynamics and withdrawal within a global land surface model, and estimated a global groundwater depletion of 330 $km^3$ $yr^{-1}$ (year 2000). The most recent study by Hanasaki *et al* (2018) estimated a non-renewable groundwater withdrawal of 182 (±26) $km^3$ $yr^{-1}$ (year 2000). Although global estimates are quite different between these studies, the regional hotspots of groundwater depletion are quite consistent with those found by Wada *et al* (2010): California's Central Valley and the High Plains aquifer in the United States, the North China Plain, western India and a part of eastern Pakistan, central Mexico, Iran and the Middle East. We refer to table 3 and figure 6 for a more extensive overview of global estimates.

Water balance methods are subject to considerable uncertainty owing to errors in biophysical, climate and socio-economic inputs and parameters of the global hydrological models used, as well as ignoring increased capture from simplified assumptions about groundwater-surface interaction (see section 2.3). Of particular interest, next to the uncertainty about global groundwater withdrawal, are the resulting large uncertainties in simulated global groundwater recharge, especially since observed recharge rates are rarely available at the scale used in global hydrological models. Natural replenishment of groundwater occurs predominantly from precipitation (i.e. diffuse recharge) and from surface water bodies such as ephemeral streams, wetlands or lakes (Scanlon *et al* 2006, Crosbie *et al* 2012, Taylor *et al* 2013). Modelled global estimates of diffuse recharge range from 11 000 to 17 000 $km^3$ $yr^{-1}$, equivalent to 30%–40% of the world's renewable freshwater resources (IGRAC GGIS, Döll and Fiedler 2008, Wada *et al* 2010, Wada and Heinrich 2013, De Graaf *et al* 2014, 2017, Hanasaki *et al* 2018). These modelled global recharge fluxes tend not to include focused recharge, which can be substantial in semi-arid environments, while preferential flow processes and the profound seasonality of recharge are equally underrepresented. Isotopes may be one way of improving recharge concepts uses in global hydrological models (Jasechko *et al* 2014). Wada and Heinrich (2013) estimated additional recharge from irrigation to be 500 $km^3$ $yr^{-1}$ globally, which is less than 5% of the global diffuse recharge, but can be substantial over arid environments.

#### *4.2.2. Volume-based methods*

The first volume-based estimate of global groundwater depletion was published by Konikow (2011). He extrapolated regional estimates, based on *in situ* groundwater level measurements, GRACE satellite observations and calibrated groundwater models, to a global estimate of 145 (±39) $km^3$ $yr^{-1}$ (average 2000–2008) using the assumption that the ratio of non-renewable to total groundwater withdrawal is spatially constant. De Graaf *et al* (2017) used a volume based method with a MODFLOW-based two-layer transient global scale groundwater model and estimated global groundwater depletion to be 7013 $km^3$ cumulatively over 1960–2010 or 137 $km^3$ $yr^{-1}$ (average). Van Dijk *et al* (2014) integrated water balance estimates derived from GRACE satellite observation, satellite water level altimetry and off-line estimates from several hydrological models, using a data-assimilation framework. The data-assimilation framework changed the estimate of global groundwater depletion derived from water balance methods from 168 to 92 $km^3$ $yr^{-1}$ (average 2003–2012). Following earlier work by Richey *et al* (2015b), Rodell *et al* (2018) recently provided a complete global assessment of terrestrial water storage trends, including groundwater depletion (figure 5(b)). Rapid advancements in large-scale hydrological modelling and increasing availability of near *in situ* satellite observation of groundwater storage change from the GRACE and its successor GRACE-FO provide a unique opportunity to better quantify groundwater depletion across the globe.

### 4.3. Regional large-scale estimates of groundwater withdrawal and groundwater depletion

As excessive groundwater withdrawal and associated depletion are highly localized, regionally parameterized and calibrated groundwater flow models (volume based

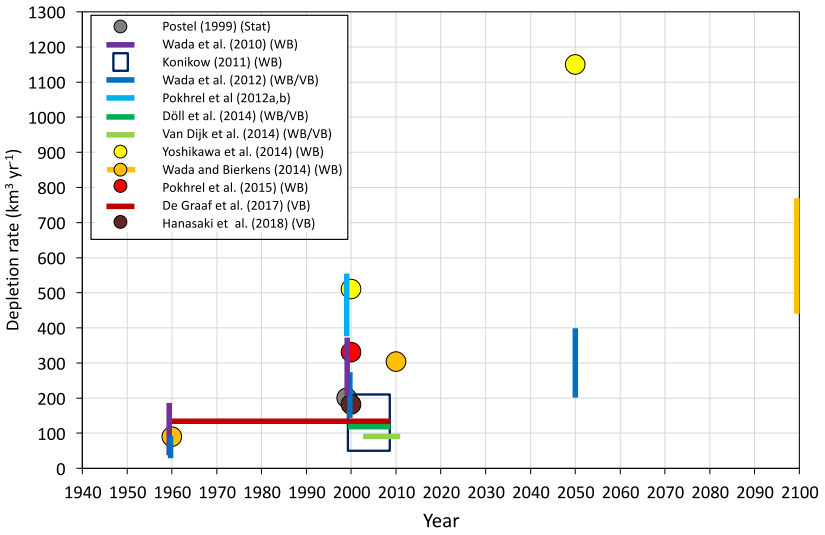


**Figure 6.** Summary of estimates of groundwater depletion; dots are single-year estimates without an uncertainty range given; horizontal lines represent temporal averages, vertical lines uncertainty bounds (estimate ±2 times the standard deviation) and boxes a combination of temporal averages and uncertainty ranges; Stat: based on reported statistics; WB: water balance method; VB: volume-based method.

methods) generally provide the better estimates of groundwater storage change (Konikow 2011, Aeschbach-Hertig and Gleeson 2012). However, in order to properly calibrate a regional groundwater model, sufficient *in situ* observations such as groundwater levels and streamflow data are needed, which are often not available, in particular in developing countries. In recent studies, GRACE-derived total terrestrial water storage changes have been increasingly applied to quantify groundwater depletion at regional scales (volume based methods). Table 4 lists recent studies that estimate groundwater depletion for various regions. These studies primarily use volume-based or partly indirect geodetic methods using regionally available data. For the North China Plain (NCP) and the California's Central Valley, both regional calibrated groundwater models and GRACE-derived approaches provide groundwater depletion estimates, while for the other regions (Northwest Sahara, Arabian, Guarani, Northern India, Bangladesh, Colorado River Basin, Canning Basin and MENA) mostly GRACE-derived estimates are available given a lack of *in situ* groundwater levels and regional groundwater models.

Scanlon *et al* (2012a) and Cao *et al* (2013) simulated the spatiotemporal variability in groundwater depletion across the North China Plain (NCP) and the two major aquifer systems in the US (California's Central Valley and High Plains Aquifer Systems) respectively, building a multilayer, heterogeneous and anisotropic flow model using MODFLOW (Harbaugh *et al* 2000). The US Geological Survey manages a dense network of groundwater level data across the country (>800 000 wells), which makes it possible to construct locally calibrated and robust groundwater models for the US. The simulated groundwater depletion estimates by Scanlon *et al* (2012a) are mostly consistent with available GRACE-derived groundwater depletion estimates (Famiglietti *et al* 2011, Scanlon *et al* 2012b). For the NCP, the groundwater depletion estimates produced with groundwater models (Cao *et al* 2013) substantially differ from those obtained using GRACE (Feng *et al* 2013, Huang *et al* 2015, Gong *et al* 2018). Huang *et al* (2015) indicated that the NCP aquifer system is highly complex, where shallow groundwater declines faster than deep groundwater, but shallow groundwater storage recovers quickly. Representing this type of complex aquifer system remains challenging even for regional groundwater models. In addition, coarse spatial resolution and noise contamination inherent in GRACE data still pose a challenge estimating groundwater depletion. Moreover, using *in situ* groundwater level observations, Shamsudduha *et al* (2012) showed that groundwater depletion estimates for the humid tropics (e.g. Bangladesh) derived from GRACE gravity estimation may be subject to large uncertainties due to highly seasonal water storage changes in other hydrological compartments.

### 4.4. Future projections of groundwater depletion

Future projections of groundwater depletion rely on hydrological model simulations that are subject to large uncertainties. Future model simulation requires climate projections from General Circulation Models or Regional Climate Models, and future socio-economic and land use change scenarios. Future land use change including agriculture is particularly important as global hotpots of groundwater depletion overlap with the areas with intensive irrigation. As land use change is heavily affected by factors such as population growth, associated food demands, economic development and international food trade, statistical extrapolation based on historical groundwater depletion

**Table 4.** Regional (large-scale) studies of groundwater depletion ($km^3\ yr^{-1}$ or $mm\ yr^{-1}$ if specified) (updated from Wada 2016).

| Region | References | Groundwater depletion $km^3\ yr^{-1}$ ($mm\ yr^{-1}$ or $Gt\ yr^{-1}$ if specified) | Year | Notes | Sources |
|---|---|---|---|---|---|
| Various regions | Sahagian *et al* (1994a) | 86.7 | Contemporary | Limited regions (e.g. USA, India, China) | Literature<br>Country statistics |
| | Famiglietti (2014) | 77.4 | 2003–2013 | Time periods vary among studies considered with limited regions | Various studies using GRACE-derived total terrestrial water storage changes |
| | Wang *et al* (2018) | 39.94 (±17.62) $Gt\ yr^{-1}$ | 2002–2016 | Endorheic basins across the globe | GRACE-derived total terrestrial water storage changes |
| Northwest Sahara | Richey *et al* (2015a) | 2.7 | 2003–2012 | Algeria, Libya, Tunisia | GRACE-derived total terrestrial water storage changes |
| Middle East and North Africa (MENA) | Foster and Loucks (2006) | 26.8 | Contemporary | | Literature<br>Country statistics |
| | Voss *et al* (2013) | 13.0 (±1.6) | 2003–2009 | Cumulative 91.3 (±10.9) $km^3$ for 2003–2009 | GRACE-derived total terrestrial water storage changes |
| Arabian | Richey *et al* (2015a) | 15.5 | 2003–2013 | Iraq, Jordan, Oman, Qatar, Saudi Arabia, UAE, Yemen | GRACE-derived total terrestrial water storage changes |
| Guarani | Richey *et al* (2015a) | 1.0 | 2003–2013 | Argentina, Brazil, Paraguay, Uruguay | GRACE-derived total terrestrial water storage changes |
| North China Plain (NCP) | Cao *et al* (2013) | 4.0 | 1960–2008 | Cumulatively 158 $km^3$ for 1960–2008 (20% of pumpage of 807 $km^3$) | Calibrated MODFLOW based groundwater model |
| | | 2.5 | 1970s | | |
| | | 4.0 | 1980s | | |
| | | 2.0 | 1990–1996 | | |
| | | 7.0 | 1997–2001 | | |
| | | 4.0 | 2002–2008 | | |
| | Feng *et al* (2013) | 8.3 (±1.1) | 2003–2010 | 2.5 $km^3\ yr^{-1}$ for shallow aquifers reported by Groundwater Bulletin of China Northern Plains | GRACE-derived total terrestrial water storage changes |
| | Huang *et al* (2015) | 2.5 (±0.4)-PP | 2003–2012 | Piedmont Plain (PP) | GRACE-derived total terrestrial water storage changes |
| | | 1.5 (±0.2)-ECP | | East Central Plain (ECP) | |
| | Gong *et al* (2018) | −17.8 (± 0.1) $mm\ yr^{-1}$-NCP | 1971–2015 | North China Plain (NCP) | Information of land subsidence, *in situ* groundwater-level measurements, literature, and GRACE satellite observations |
| | | −76.1 (±6.5) $mm\ yr^{-1}$-B | 1999–2012 | Beijing (B) | |

**Table 4.** (Continued.)

| Region | References | Groundwater depletion $km^3\ yr^{-1}$ ($mm\ yr^{-1}$ or $Gt\ yr^{-1}$ if specified) | Year | Notes | Sources |
|---|---|---|---|---|---|
| Indus | Cheema *et al* (2014) | 31 | 2007 | 68 $km^3$ of total groundwater abstraction | Remote sensing combined with a hydrological model and spatial information on canal water supplies |
| Northern India | Rodell *et al* (2009) | 17.7 (±4.5) | 2002–2008 | Cumulative 109 $km^3$ for 2002–2008 | GRACE-derived total terrestrial water storage changes |
| Northern India and surrounding regions | Tiwari *et al* (2009) | 54 (±9) | 2002–2008 | | GRACE-derived total terrestrial water storage changes |
| | Jacob *et al* (2012) | 35 | 2003–2010 | | GRACE-derived total terrestrial water storage changes |
| Bangladesh | Shamsudduha *et al* (2012) | 0.44 (±1.24)-2.04 (±0.79)-wet seasons | 2003–2007 | Depletion of 0.52 (±0.30)-0.85 (±0.17) $km^3\ yr^{-1}$ from borehole hydrographs | GRACE-derived total terrestrial water storage changes |
| | | 0.52(±0.5)-2.83(±0.42)-annual | | | |
| California's Central Valley | Famiglietti *et al* (2011) | 3.1 (±0.6) | 2003–2010 | Cumulative 20.3 $km^3$ for 2003–2010 | GRACE-derived total terrestrial water storage changes |
| | Scanlon *et al* (2012a) | 2.0 | 1962–2003 | Cumulative 24.6 $km^3$ for 1976–1977, 49.3 $km^3$ for 1987–1992, 140 km3 since the 1860s, and 80 $km^3$ since the 1960s | Calibrated MODFLOW based groundwater model |
| | | 6–8 | 2006–2010 | | |
| | Scanlon *et al* (2012b) | 8.9 (±0.9) | 2006–2010 | Cumulative 31.0 (±3.0) $km^3$ for 2006–2010 | GRACE-derived total terrestrial water storage changes |
| High Plains Aquifer | Scanlon *et al* (2012a) | 5.7 | 1950–2007 | Cumulative 330 $km^3$ after pre-development in the 1950s | Calibrated MODFLOW based groundwater model |
| | | 7.0 | 1987–2007 | | |
| | | 12.5 | 2003–2013 | | |
| Colorado River Basin | Castle *et al* (2014) | 5.6 (±0.4) | 2004–2013 | Cumulative 50.1 $km^3$ groundwater loss out of 64.8 $km^3$ freshwater loss | GRACE-derived total terrestrial water storage changes |
| Canning Basin | Richey *et al* (2015a) | 3.6 | 2003–2013 | Australia | GRACE-derived total terrestrial water storage changes |

rates is likely not suitable to project future groundwater depletion. Thus far, only few studies are available that attempt to assess future groundwater depletion (table 2). The first global study was produced by Wada *et al* (2012b) who projected future groundwater depletion based on the combination of three climate and socio-economic scenarios. Their results showed that global groundwater depletion is projected to increase from 204 (±30) km$^3$ yr$^{-1}$ in 2000 to 295 (±47) km$^3$ yr$^{-1}$ by 2050. Yoshikawa *et al* (2014) found a much higher depletion volume under a consistent expansion of irrigated areas and projected global groundwater depletion to reach ~1150 km$^3$ yr$^{-1}$ by 2050. They, however, used a globally medium population growth scenario (0.9% yr$^{-1}$) to extrapolate the future irrigated area change, which is rather high (from 2.7 in 2000 to 3.9 million km$^2$ in 2050), as the expansion of irrigated areas has been slowing down in many countries. Wada and Bierkens (2014) quantified the fraction of the consumptive blue water use that is met from non-sustainable use of groundwater and surface water. They projected global total and non-renewable groundwater withdrawal and depletion to increase from 952 and 304 km$^3$ yr$^{-1}$ in 2010 to respectively 1621 (±128) and 597 (±85) by the end of this century. For climate and socio-economic change they used a business as usual scenario based on the latest Representative Concentration Pathways (RCP6.0) and Shared Socioeconomic Pathways (SSP2). These future projections show that, apart from an intensification of a number of current hotspots, groundwater depletion is likely to expand to other regions such as Africa, other parts of Asia, and South America where significant increase in population (>2 billion), food production and economic development are expected. In developed economies such as the US, the annual rate of groundwater depletion in the High Plains Aquifer is estimated to have already peaked at 8.25 km$^3$ yr$^{-1}$ in 2006 followed by projected decreases to 4 km$^3$ yr$^{-1}$ in 2110 (Steward and Allen 2016). This indicates that we can expect a rapid increase in groundwater depletion in developing economies and gradual decrease in developed economies such as the US. We stress, however, that projecting future demand is notoriously difficult leading to large differences between projections and models (Wada *et al* 2016a) and improvements are desperately needed. This would mean going beyond the current practice of using relatively simple statistical modelling of relationships between socio-economic data and water withdrawal data, possibly using more advanced behavioural modelling based on machine learning and agent-based modelling (Aerts *et al* 2018, Mason *et al* 2018).

### 4.5. Estimates of global groundwater volumes

The perpetual use of non-renewable groundwater will eventually result in complete exhaustion of groundwater stocks. This raises the question: 'How long will it last?'. In order two answer this question two underlying questions need to be answered: (1) 'How much groundwater is there?' and (2) 'What will be the future withdrawal rates?'. As groundwater is not a global common pool, these questions will result in different answers depending on region and even location. As has been shown in previous studies (Gleeson *et al* 2015, Richey *et al* 2015a), there is enormous uncertainty in estimates of global groundwater stocks as well as in regional assessments. Moreover, scenarios about future water demand are still under development, while projections under the same scenario are extremely variable between global models (Wada *et al* 2016a). This makes the question 'How long does it last' a wicked problem. In this review, we will focus on the first question only.

According to Shiklomanov (1993) groundwater encompasses 30% of total fresh water storage and over 98% of the non-frozen fresh water storage on land. Given the limited knowledge about the subsoil (Bierkens 2015), estimates of total groundwater volume are, however, extremely uncertain as testified by the different estimates shown in table 5 that range between 1 and 60 km$^3$ globally. Recent studies (Gleeson *et al* 2015, Richey *et al* 2015a) have reviewed extensively earlier estimates of global groundwater volume. Our review is largely based on these assessments. Generally, all estimates rely on the assumption that the free pore space below the water table is filled with water and that porosity diminishes according to a certain function with depth (either by layer or exponentially). The general equation to estimate the volume (with $A$ total aquifer area, $n$ porosity, $z$ depth, $d_{gw}$ depth to groundwater and $\mathbf{x}$ location within an aquifer):

$$V = \int_A \int_{d_{gw}}^{\infty} n(\mathbf{x}, z)\,\mathrm{d}z\,\mathrm{d}\mathbf{x}. \quad (3)$$

The oldest estimate (Vernadskiy 1933) is also the largest and equals 60 million km$^3$; however, this number covers depths to 5 km at which it is highly unlikely that any extractable fresh groundwater can be found and is therefore considered to be on the high side. Another earlier estimate is from Nace (1969). This estimate (7 million km$^3$) has long been assumed to be on the low side because of the low porosity assumed (1%). Also, the arbitrary multiplication with a factor 5 makes this estimate questionable. The estimate of Korzun (1978) (23.4 million km$^3$) is generally seen as the most acceptable one as testified by its appearance in multiple text books on global water resources (e.g. Shiklomanov 1993). Recently, using a much more data-intensive approach, Gleeson *et al* (2015) arrived at a surprisingly similar estimate, which should increase confidence in Korzun's textbook estimate. At the same time Gleeson *et al* (2015) state that not all of this groundwater will be of sufficient quality to be useable. Moreover, because porosity and permeability are low at large depths, it would most likely be technically infeasible to extract the groundwater in sufficient quantities to be of use. Following this reasoning,

**Table 5.** Estimates of global aquifer storage (extended from an inventory by Gleeson *et al* 2015, Richey *et al* 2015a).

| Extent, depth/domain and detail | Global estimate | Method | References |
|---|---|---|---|
| Global, 1000 m | $7 \times 10^6$ km$^3$ | Assuming single effective porosity of 1% up to 1000 m depth times 5 | Nace (1969) |
| Global, 2000 m, per continent | $23.4 \times 10^6$ km$^3$ | Total volume: Three layers 100, 200, 200 m thickness with effective porosity of 15%, 12%, 5% | Korzun (1978) |
| | $10.5 \times 10^6$ km$^3$ (fresh only) | | Shiklomanov (1993) |
| Global, 5000 m | $60 \times 10^6$ km$^3$ | Citing previous research of Vernadskiy (1933) and Makarenko (1966) based on the non-chemically bounded water in the upper crust | L'vovich (1979) |
| Global, saturated thickness of 200 m, Major aquifers of the world | 1.1 (0.6–1.6) $\times 10^6$ km$^3$ | Fixed minimum and maximum specific yield per aquifer based on major aquifer lithology and using a saturated thickness of 200 m. | Richey *et al* (2015a) |
| Global, 2000 m, | Total: 22.6 (15.8–29.5) $\times 10^6$ km$^3$ | Total: volume fraction 4 lithology classes and depth-porosity profiles per class | Gleeson *et al* (2015) |
| Both total groundwater volume and modern <50 m groundwater volume (over 900 000 watersheds) | Modern (<50 years): 0.35 (0.24–3.8) $\times 10^6$ km$^3$ | Modern: groundwater modelling and tritium dating | |
| Global at 5 arcminutes, only sedimentary deposits and sedimentary rocks in major aquifers | $6 \times 10^6$ km$^3$ | Based on mapped aquifer thickness and porosities (fixed with depth) in de PCR-GLOBWB/MODFLOW global groundwater model | De Graaf (2016) |

Richey *et al* (2015a) arrived at much lower volumes of extractable groundwater (1.1 million km$^3$) assuming that for most aquifers the saturated thickness from which groundwater can be withdrawn is on average 200 m (following Margat and Van der Gun 2013). Their results match considerably better with volume estimates from regional aquifer studies. This, however, negates that sometimes multiple aquifers systems are stacked on top of each other that collectively may amount to much more extractable groundwater. Moreover, regional studies generally focus on the shallower systems that are more easily exploited and sampled. Based on the hydrogeological parameterization of a global groundwater model (i.e. De Graaf *et al* 2015, 2017), De Graaf (2016) provides an intermediate estimate of groundwater in sedimentary basins of 6 million km$^3$.

The current consensus seems to be that the global groundwater storage amounts to approximately 23 million km$^3$. However, the physical limit that matters is the extractable and useable volume. The useable volume depends on the quality (mostly salinity) of the groundwater, with the fresh groundwater volume being less than half of the total according to Shiklomanov (1993). The extractable volume depends on four important parameters: (1) the depth at which the groundwater should be pumped, i.e. how deep the filter and pump should be installed; (2) the static head in the well compared to the surface level which determines the amount of lift that is needed; (3) the drainable porosity or storage coefficient of the layer from which groundwater is pumped which determines the volume extracted per m head decline; (4) the permeability of the aquifer which determines the maximum yield (m$^3$ s$^{-1}$) that can be achieved by the well, which should be sufficiently high to support the water demand by e.g. irrigation. It should therefore be concluded that in order to assess globally and regionally extractable and useable groundwater volumes, equation (3) will not suffice and a global groundwater flow and transport model is needed. At the time of writing this review, there is only one model available that can simulate global groundwater flow and withdrawal (De Graaf *et al* 2017). The hydrological schematization underlying this model is, however, quite rudimentary and needs improvement. Table 6 provides an overview of global datasets that are currently available to parameterize global groundwater flow and transport models. Again, they are not yet sufficiently detailed for accurate regional and global estimates and would require updating using smart combinations of geological maps, regional groundwater model studies, data from observation wells and bore logs.

We conclude this section by stating that the parameters that determine the volume of extractable groundwater may be only partly of a technical nature, since they are also subject to economic laws. Similar to oil, the technical efforts suffered to extract groundwater highly depend on the economic value of the water when used (Burt 1964, Gisser and Sanchez 1980). If high enough, an increasing economic value of

**Table 6.** Global hydrogeological datasets.

| Dataset type and name | Method of creation | Resolution | References |
|---|---|---|---|
| *Aquifer extent and type* WHYMAP | Assemblage from geological and hydrogeological maps | Vector map 1:25 000 000 | Richts *et al* (2011) |
| *Surface lithology* GLIMS | Assemblage from national geological maps | vector map 1235 400 polygons | Hartmann and Moosdorf (2012) |
| *Permeability and porosity* GLHYMPS | Combining surface lithology maps with data from US groundwater studies (<50 m) | Version 1: vector map: average polygon size 14 000 km$^2$ | Gleeson *et al* (2011) and |
| | | Version 2: vector map: average polygon size 100 km$^2$ | Gleeson *et al* (2014) |
| *Depth to Bedrock including soil, regolith layers and unconsolidated sedimentary deposits* | | | |
| <50 m | From geomorphological mapping and assessing class-specific relations using US borehole data | Grids 30 × 30 arc-seconds | Pelletier *et al* (2016) |
| All depths | From machine learning mapping using various global datasets including boreholes | Grid 250 × 250 m | Shangguan *et al* (2017) |
| All depths, only coastal aquifers | Combined topographical and Geology-based mapping using different datasets (GEBCO, De Graaf *et al* 2015, Pelletier *et al* 2016) | Transects every 5 km across the global coast | Zamrsky *et al* (2018) |
| *Depth to Bedrock including unconsolidated sedimentary deposits and sedimentary rocks (permeable part)* | | | |
| | Mapped from seismic data | Grids 1 × 1 arc-degrees | Limberger *et al* (2018) based on Laske and Masters (1997) and Tesauro *et al* (2008). |
| PCR-GLOBWB/MODFLOW | Statistical relations to topography and calibrating these with US groundwater modelling studies | Grids 30 × 30 arc-seconds | De Graaf *et al* (2015) |

water will spark innovations that would enlarge the volume of usable groundwater that can be extracted.

## 5. Impacts of groundwater withdrawal and groundwater depletion

Groundwater withdrawal and groundwater depletion result in several adverse impacts (see e.g. Custodio 2002, Dumont 2013 for overviews). Here we will review a few examples of these.

### 5.1. Land subsidence

In section 3 we have already introduced the effects of groundwater depletion on land subsidence, as it makes it possible to monitor depletion using geodetic observations. Land subsidence has quite a number of detrimental effects, such as damage to infrastructure (underground pipes, roads, bridges, building) and increased flood risk by subsidence-related dike failure and lower surface levels (Sato *et al* 2006). We will not repeat what was written in section 3, but instead will add some more information about the processes behind land subsidence by groundwater withdrawal. Subsidence occurs because a release of pore pressure causes water carrying layers to be compacted under the weight of the overlying sediments, or because the water table falls below clay or peat layers, which results in shrinking of these materials. Part of the land subsidence is elastic. This means that if groundwater heads and/or water tables would be restored to the old values, the land would rise again. There is also a part that is inelastic, either by irreversible shrink, mineralization and oxidation of organic materials, or by a rearrangement/resettling of grains (Galloway *et al* 1998, van Asselen *et al* 2009). The problem with inelastic subsidence is not only that land elevation cannot be restored to its original value, but also that the storage capacity of the aquifer is compromised, which means that it is not possible to recover full storage after groundwater withdrawal stops. Inelastic subsidence is often considered as plastic instantaneous deformation (e.g. Leake and Galloway 2007) which may be appropriate for sandy deposits. If, however, clay and peat layers are present (as in many subsiding regions), inelastic compaction is time dependent and results in secondary compaction or creep (Minderhoud *et al* 2017). Creep is believed to be the cause of

prolonged subsidence, even when withdrawal has stopped. For instance, when the city of Tokyo drastically reduced groundwater withdrawal rates to cease subsidence, subsidence still continued for years, albeit at a much slower rate (Sato *et al* 2006).

### 5.2. Enhancement of hydrological drought

A meteorological drought is defined as a prolonged period with below-normal precipitation. (Tallaksen and van Lanen 2004). If persistent enough, a meteorological drought will propagate through the hydrological system by causing abnormally low soil moisture contents and evaporation (agricultural drought), reduced groundwater recharge and low water tables (groundwater drought) and finally declining groundwater discharge to streams resulting in extremely low streamflow (hydrological drought) (Wilhite and Glantz 1985, Tallaksen and van Lanen 2004). As excessive groundwater withdrawal affects groundwater discharge, it may therefore also enhance hydrological drought (Wada *et al* 2013). The Ganges River Basin, home to over 400 million people, includes the aquifers with highest groundwater depletion rates around the word (Gleeson *et al* 2012). In recent years, the Ganges Basin is experiencing widespread reduction of summer streamflow. A recent study by Mukherjee *et al* (2018) reports that severe groundwater depletion is the likely cause of this summer flow drying due to decreasing groundwater contribution to streamflow over the region. Decreasing summer flows trigger other environmental problems, such as degrading water quality due to reduced dilution of pollutants and pathogens and higher water temperatures. Also, the subsequent reduction of surface water flows in turn lead to potential water deficits for agricultural production downstream, resulting in further groundwater depletion over large agricultural regions (Scanlon *et al* 2012a, 2012b). This increased groundwater withdrawal compensates, albeit temporarily, for decreased surface water availability (Taylor *et al* 2013).

### 5.3. Contribution to sea-level change

Another consequence of groundwater depletion is the contribution to global sea level rise. Deep fossil groundwater has been isolated from the current hydrological, atmospheric, and ocean balance and cycle for hundreds to thousands of years, depending on storage volume, recharge, and discharge rates (see section 2.2). Withdrawal of this groundwater will thus redistribute water stored on land to the hydrological cycle and contributes to additional ocean storage and sea-level rise.

Sahagian *et al* (1994a) were the first to try to estimate the contribution of terrestrial water storage change to global sea-level variation. The main components of this positive and negative contribution are impoundments behind reservoirs (negative), groundwater depletion (positive), wetland loss (positive) and changes of water levels in lakes and endorheic basins (negative or positive). Subsequent estimates of the sign and magnitude of this terrestrial component differ greatly (Chao 1994, Gornitz *et al* 1994, Greuell 1994, Rodenburg 1994, Sahagian *et al* 1994b) and have been subject to considerable debate. This resulted in the IPCC fourth assessment report (IPCC 2007) neglecting the contribution of non-frozen terrestrial waters to sea-level variation, due to its perceived uncertainty and the assumption that negative contributions such as impoundments behind dams compensate for positive contributions such as groundwater depletion. Subsequent studies (Postel 1999, Gornitz 2000, Huntington 2008, Milly *et al* 2010, Church *et al* 2011) differ mostly in their estimates of the contribution of groundwater depletion, due to methodological differences, i.e. based on limited country estimates versus based on global hydrological modelling.

More recent work on global groundwater depletion (Wada *et al* 2010, Konikow 2011, Wada *et al* 2012b) suggests an increase of a positive contribution to sea-level rise during the last decade as a result of a rise in groundwater depletion. For instance, Wada *et al* (2012b) project an increase of the contribution of global groundwater depletion to sea level rise from 0.57 ($\pm$0.09) mm yr$^{-1}$ in 2000 to 0.82 ($\pm$0.13) mm yr$^{-1}$ towards 2050. The increase is primarily driven by growing water demand during the last century, but is also affected by decreased water availability and groundwater recharge, and larger evaporative demand from agricultural areas due to changes in precipitation patterns and higher temperatures. Other studies report the contribution of groundwater depletion to global sea level rise in the order of 0.3–0.9 mm yr$^{-1}$ for around the year 2000 (Wada *et al* 2010, Konikow 2011, Döll *et al* 2014, Pokhrel *et al* 2015, De Graaf *et al* 2017, Hanasaki *et al* 2018). In response to these more recent studies, the terrestrial contribution to sea-level change was again included in the sea-level chapter of the fifth IPCC report (IPCC 2013).

The aforementioned estimates assume that nearly 100% of groundwater extracted eventually ends up in the oceans. A recent study by Wada *et al* (2016) used a coupled climate-hydrological model simulation to show that only 80% of groundwater depletion ends up in the ocean, while the rest is recycled by local precipitation. The resulted contribution of groundwater depletion to global sea level rise eventually amounts to 0.02 ($\pm$0.004) mm yr$^{-1}$ in 1900 and increased to 0.27 ($\pm$0.04) mm yr$^{-1}$ in 2000, which indicates that existing studies have substantially overestimated the contribution (Wada *et al* 2016b). To add to this, a study by Reager *et al* (2016) used more than 12 years of GRACE gravity estimates to show that over the period 2002–2014 the positive contribution of groundwater depletion of 0.38 mm yr$^{-1}$ was more than offset by a negative contribution of water stored on land ($-$0.71 mm yr$^{-1}$) as a result of increased precipitation due to climate variability or change. Wang *et al* (2018),

however, published estimates of a strong positive contribution (0.295 mm $yr^{-1}$) from water loss from endorheic basins of which about 40% was attributed to groundwater. The work of Wada *et al* (2016), Reager *et al* (2016) and Wang *et al* (2018) show that the issue of the contribution of groundwater depletion to sea-level rise is far from resolved.

### 5.4. Groundwater salinization

In the top 100–500 m of the of larger aquifer systems we can expect groundwater to be fresh (concentration of total dissolved solids TDS smaller than 1 g $l^{-1}$). Below these depths groundwater is most likely brackish (TDS 1–10 mg $l^{-1}$), saline (TDS 10–35 mg $l^{-1}$) or hyper-saline (> 35 mg $l^{-1}$), where the origin of natural high salinity groundwater can be either marine or terrestrial (Van Weert *et al* 2009, Margat and Van der Gun 2013, Van Engelen *et al* 2018). In some aquifers or regions, brackish or saline groundwater can occur at shallow depths. In these regions, groundwater pumping, due to a decrease in groundwater pressure below the pumping well screen, can lead to upconing of the fresh-salt groundwater interface resulting in salinization of groundwater resources. Van Weert *et al* (2009) provide a global inventory of regions where brackish and saline groundwater can be found at intermediate or shallow depths.

The pumping of groundwater has led to the progressive salinization of groundwater in many parts of the world, particularly in coastal aquifers, as testified by a large number of case studies described in e.g. Custodio and Bruggeman (1987) and Post *et al* (2018). The process of upconing of a salt-fresh groundwater interface under an individual wells or drains has been studied extensively. Earlier work (Bear and Dagan 1964, Dagan and Bear 1968, Schmorak and Mercado 1969, Strack 1976) and subsequent extensions (e.g. Reilly and Goodman 1987, Garabedian 2013) provide analytical solutions to the upconing elevation of the fresh-salt groundwater interface and the critical pumping rates that lead to salinization of the well. These studies show that, if the fresh-salt water interface is not too far below the well screen, upconing and well contamination occurs rather quickly (order of years). They also show that return of the salt water cone to its original state may take an order of magnitude longer (decades), especially when groundwater recharge is small.

These analytical studies all assume a sharp boundary between fresh and salt groundwater, which generally results in optimistic estimates of the effects of pumping on salinization. If hydrodynamic dispersion is taken into account, a brackish transition zone will develop that has a larger area of influence of brackish upconing then a sharp interface (Reilly and Goodman 1987, Zhou *et al* 2005, Jakovovic *et al* 2016) and creates a gradual increase in concentrations of pumped groundwater (Reilly and Goodman 1987, Werner *et al* 2009). Also, the development of a brackish transition zone makes the decay of salinity after pumping has stopped much slower. Thus, numerical models including hydrodynamic dispersion and experimental results indicate that in case salt groundwater is present closely below the well screen, groundwater withdrawal may not only result in well shutdown within a short period of time, but will also render groundwater unsuitable for use over a large area around the well (several $km^2$, Jakovovic *et al* 2016) for prolonged periods (several decades, Zhou *et al* 2005).

In conclusion, areas with intensive groundwater withdrawal and shallow saline groundwater (Van Weert *et al* 2009), most prominently coastal aquifers, are very sensitive to irreversible salinization of fresh groundwater resources, which is likely to become a global problem (Custodio 2002, Konikow and Kendy 2005). This is further supported by a sensitivity study by Ferguson and Gleeson (2012), who find that the impact of groundwater withdrawal on coastal aquifers is more significant than the impact of sea-level rise for a wide range of hydrogeologic conditions and withdrawal intensities. Despite the global importance of salinization by groundwater withdrawal and reported cases of imminent salinization (Custodio and Bruggeman 1987, Post *et al* 2018), there are hardly any regional-scale projections of aquifer salinization under future developments of groundwater withdrawal, with Mabrouk *et al* (2018) as a recent exception.

### 5.5. Impact on groundwater-dependent ecosystems

Groundwater withdrawal has impact on streamflow, groundwater levels and evaporation (figure 4) and as such on groundwater dependent ecosystems (GDEs), i.e. ecosystems with organisms that depend on groundwater discharge or the proximity of a water table. Eamus *et al* (2015, 2016) divides GDEs into three classes: (1) GDEs that reside within groundwater itself (e.g. stygofauna in caves, fissures); (2) GDEs requiring the surface expression of groundwater (springs, lakes, streams, wetlands) and (3) GDEs dependent upon the availability of groundwater within the rooting depth of vegetation (e.g. woodlands; riparian forests). In earlier work, Foster *et al* (2006) used a classification based on their geomorphological setting (aquatic, terrestrial, coastal) and associated groundwater flow mechanism (deep or shallow). Their insightful figure 1 is reproduced here (figure 7) and shows that classes A–D fit into category 2 of Eamus *et al* (2015) and class E into category 3. Recent overview papers about global GDEs are given by Eamus *et al* (2015) (monitoring) and Rohde *et al* (2017) (policy and management). Doody *et al* (2017) provide a framework for continental-scale mapping of GDEs using remote sensing and expert knowledge. Of interest to mapping GDEs is the global groundwater depth map provided by Fan *et al* (2013) based on a 30 arc-second steady state global

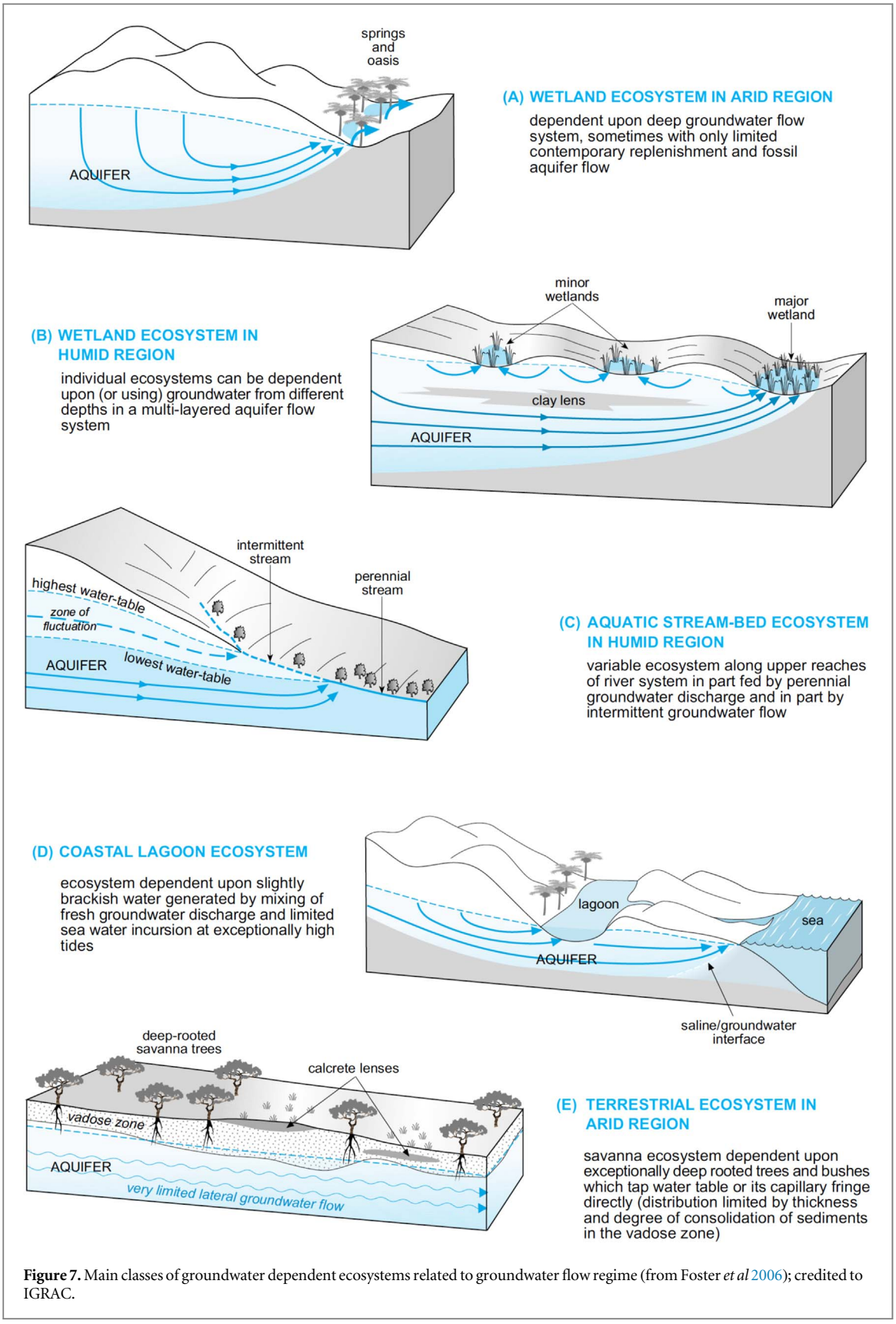


**Figure 7.** Main classes of groundwater dependent ecosystems related to groundwater flow regime (from Foster *et al* 2006); credited to IGRAC.

groundwater model. From this map Fan *et al* (2013) estimate that ~15% of the land surface (not including the large lakes) is covered by areas that receive persistent groundwater discharge (lakes, marshes, swamps, fens, springs, streams), ~2% by less frequently inundated wetlands (floodplains an fens) and 5%–15% with the water table depth within the rooting depth of upland plants.

From figures 4 and 6 we can deduce that for cases A–D the effect of groundwater withdrawal on GDEs is most prominent during stage 1 withdrawal, while both stage 1 and stage 2 withdrawal affect case E. When

discussing the effect of groundwater withdrawal on GDEs we will distinguish between aquatic ecosystems (C and D in figure 7) and terrestrial ecosystems (A, B and E in figure 7). It should, however, be noted that this subdivision is partly arbitrary as GDEs such as marshes and swamps possess elements of both.

*5.5.1. Aquatic ecosystems*

Most of the literature about the impact of water use on aquatic ecosystems is focused on environmental flow limits. The term 'environmental flow' refers to the quantity, quality and timing of water flows to sustain freshwater and estuarine ecosystems and the human livelihoods that depend on these (The Brisbane Declaration 2007). There exists a huge body of literature on environmental flows and quite a number of reviews exist (e.g. Olden and Poff 2003, Tharme 2003, Linnansaari *et al* 2013, Pastor *et al* 2014) with over 200 different environmental flow methods in use today. Many of the methods have only been used locally or regionally and it is difficult to choose which one is applicable at larger scales. Tharme (2003) and subsequent reviews distinguish between: hydrological methods, which characterize flow alteration with different flow characteristics before and after human impact (Tennant 1976, Smakhtin *et al* 2006), Hydraulic methods (e.g. Reiser *et al* 1989) that additionally identify hydraulic parameters such as water depth, wetted parameter and flow velocity; Habitat simulation, whereby ecological data such as species abundance are correlated with flow and temperature characteristics; Holistic methods (e.g. ELOHA, Poff *et al* 2010) in which hydrologic, hydraulic and habitat simulation methods are combined.

In the ELOHA framework, the importance of groundwater discharge for sustaining minimum flows through low flow conditions is recognized. Nevertheless, the direct effect of groundwater withdrawal on environmental flows has only been published on recently (Barlow and Leake 2012, Acreman *et al* 2014, Hendriks *et al* 2014, Gleeson and Richter 2018). The latter paper is of particular interest because it tries to define a presumptive standard on groundwater-to-stream discharge itself, which could be used, together with a simple groundwater parameterization, as an environmental limit to groundwater withdrawal. Looking at figure 4, we note that the effect of groundwater withdrawal on streamflow is mainly present in stage 1 withdrawal (figures 4(b) and (c)). In case the water table becomes disconnected from the stream (stage 2) further increase of withdrawal rates will lead to increased depletion, but will barely impact environmental flows any further. Finally, it should be noted that even if groundwater withdrawal occurs from underneath a confining layer, i.e. from a confined aquifer, the environmental flows from the larger streams that (almost) penetrate this layer can still be affected (Barlow and Leake 2012, Hendriks *et al* 2014).

Streams that are heavily influenced by groundwater discharge generally have lower stream water temperatures during low flow periods (Stark *et al* 1994, Risley *et al* 2010). Moreover, groundwater discharge is often nutrient poor and of different pH than streamflow which not only impacts stream water chemistry directly (Caissie *et al* 1996, Sear *et al* 1999), but also indirectly through biogeochemical processes associated with mixing of groundwater and surface water in the hyporheic zone (Merill and Tonjes 2014). Thus, apart from affecting ecosystems by altering flow regimes, groundwater withdrawal also affects aquatic ecosystems by altering stream water quality. However, due to the complexity of the relationship between groundwater discharge and stream water quality, no attempts have been made to derive the associated environmental limits to groundwater withdrawal.

*5.5.2. Terrestrial ecosystems*

Figure 4 also shows that the lowering of the phreatic surface by groundwater withdrawal impacts riparian vegetation through reduced evaporation (e.g. Shafroth *et al* 2000). Apart from the direct individual effects on evaporation and therefore productivity of phreatophytes (Yin *et al* 2018), the lowering of water tables also affects the competitive advantages of rare species that have been adapted to shallow water tables and oxygen-poor soils (Runhaar *et al* 1997, Elmore *et al* 2006). For instance, Runhaar *et al* (1997) provide strong relationships between the relative abundance of hydrophytes and xerophytes and mean spring water table depth. Also, evident from this work is these relationships are only strong if they are obtained separately per soil type. The physical explanation for this is the different degrees of soil aeration between soil types for a given water table depth.

The lowering of the phreatic surface also has an impact on the soil water chemistry. Areas of shallow water tables are often dominated by nutrient poor alkaline rich groundwater exfiltration. These circumstances benefit low productive fen ecosystems harbouring rare species (Witte *et al* 2015). A drop of the water table sets in motion a chain of effects that heavily influences species composition (Lamers *et al* 2002): first, desiccation itself generates acidification through the chemical oxidation of iron and sulphide, yielding sulphuric acids. This effect, together with the replacement of alkaline groundwater with infiltrating rainwater, reduces the pH of soil water drastically, which leads to an increased mobility of potentially toxic metals, including heavy metals. Also, increased oxygen availability resulting from lower water tables induces mineralization of organic matter in the soil which increases nutrient availability, in particular nitrogen (Grootjans *et al* 1986). As phosphorous is bounded more strongly to soils in dry circumstances due to iron-oxidation, the nitrogen to phosphorous ratio (N/P) changes as well. Thus, the lowering of groundwater levels, i.e. by agricultural drainage or

groundwater withdrawal, not only affects abundance of riparian vegetation and phreatophytes by desiccation, but also the species composition of low-productive fen ecosystems through soil-groundwater biogeochemistry (van Loon *et al* 2009). Although it has not been done yet, linking groundwater withdrawal limits to water table levels and associated plant evaporation reduction seems feasible. Determining withdrawal limits related to groundwater-related changes in ecosystem biogeochemistry is, however, far more complex and remains a huge challenge.

## 6. Hydroeconomics of groundwater use

In its most simple form, dealing with the economics of groundwater withdrawal means confronting the benefits from groundwater exploitation with the costs of withdrawal. Economists tend to take a much wider view on the exploitation of natural resources, including groundwater, where concepts such as profit maximization, opportunity costs (also across time), externalities and the relationship between demand/supply and price play a role. Within the context of this review, it would be impossible to provide an overview of the enormous body of economic literature dealing with the economics of natural resources. We will refer to handbooks and review collections on the subject (e.g. Halvorsen and Layton 2015, Halvorsen 2018). Limiting ourselves to hydroeconomics and in particular hydroeconomic modelling, reviews can be found in Brouwer and Hofkes (2008), Harou *et al* (2009) and more recently Bekchanov *et al* (2015) and and Bauer-Gottwein *et al* (2017). From these reviews we find that, amongst others, hydroeconomic models are used for simulation and optimisation of water use and allocation within river basins (Houk *et al* 2007, Pulido-Velazquez *et al* 2008), analysing the economy-wide effects of water scarcity (Konar *et al* 2013, Lenzen *et al* 2013), the role of non-renewable groundwater in food trade (Dalin *et al* 2017), the guidance of investments in water infrastructure (Rosenberg *et al* 2008), analysing the effectiveness of taxes and water pricing (Höglund 1999, Medellín-Azuara *et al* 2015, Macian-Sorribes *et al* 2015, Rougé *et al* 2018) and water markets (Cummings and Nercissiantz 1992, Characklis *et al* 2006) in promoting optimal or sustainable water use, and the valuation and design of adaptation measures to climate change impacts (Escriva-Bou *et al* 2017, Girard *et al* 2015). In this section which is focused on the hydroeconomics of groundwater use, we will first examine the special properties of groundwater when viewed as a natural resource. Next, we will review the hydroeconomic literature on assessing optimal groundwater withdrawal rates. We end with listing a number of economic and voluntary incentives to make groundwater use more sustainable.

### 6.1. Properties of groundwater as a natural resource

Groundwater as a natural resource has a number of properties that make it special. Some important ones are:

(1) Ownership is often poorly established or groundwater is jointly owned by the land owners sitting on top of the aquifer (Moench 1992, Theesfeld 2010). This entails that without a permit system in place, anyone with the capital to install a production well on his/her land is allowed to use it (Famiglietti 2014), although legally established limits may apply.

(2) Groundwater can be both a renewable as well as a non-renewable resource, which calls for a different economic analysis (Halvorsen and Layton 2015).

(3) Extraction (i.e. withdrawal) costs are not constant per unit extracted as assumed in many economic analyses and increase with groundwater depth (e.g. Gisser and Sánchez 1980, Negri 1989, Foster *et al* 2015).

(4) The benefits of groundwater withdrawal are for the individual user, while the costs of the exploitation are (partly) borne by all the users of the same aquifer system. This occurs because groundwater withdrawal may lead to water table (or head) decline in parts of the aquifer that do not underlie the individual user's land, thus leading to increased extraction costs for others, often called pumping externalities (Negri 1989). So, just as with many other common pool resources groundwater may suffer from the tragedy of the commons (Hardin 1968, Müller *et al* 2017). Note, however, that the largest groundwater decline will occur at the location where groundwater is being pumped.

(5) Groundwater is often not exactly a common pool resource. First of all, it is not a fully non-exclusive resource (Gisser and Sánchez 1980), because generally only the owners of land overlying the aquifers can access the groundwater it contains. Second, the degree to which groundwater can be seen as a common pool depends the hydrogeological properties of the aquifer. For instance, groundwater withdrawal from a homogenous sedimentary confined aquifer with large transmissivity results in a large radius of influence of the pumping wells. Such an aquifer thus resembles a common pool. On the other hand, a granite aquifer with isolated pockets of high secondary permeability cannot be seen as a common pool.

(6) Being (partly) a common pool resource entails that the costs of groundwater pumping surpass the direct withdrawal costs (i.e. the well construction costs and energy costs to lift water above the

ground). Rogers *et al* (1998) list these additional costs that, when not accounted for, hamper the efficient use of groundwater: opportunity costs resulting from depriving other more profitable types of water use (now and in the future), environmental externalities, such as the costs of ecosystem deterioration due to lowering water tables and diminished low flows; and economic externalities, for instance increased extraction costs for future users due to groundwater level decline or wells drying up and becoming stranded assets (Perrone and Jasechko 2017).

(7) Another property that arises from the common pool characteristic of groundwater is so-called 'strategic externality' (Negri 1989). This means that what a groundwater user does not extract today, is likely to be extracted by a rival user today or tomorrow. This property frustrates any incentives to forego current pumping and leave groundwater in the ground for future use.

Economic theories that seek to optimize groundwater use should take account of these special properties of groundwater resources. Hereafter, we provide an overview of past research on optimal groundwater withdrawal and depletion.

### 6.2. Optimal groundwater withdrawal and depletion

When examining approaches used to analyse optimal groundwater withdrawal, one can distinguish between papers by economists that use sophisticated mathematical analyses but relatively simple aquifer representations, e.g. bathtub models or single cell aquifers, and papers by hydrogeologists that use more realistic numerical models of groundwater flow and withdrawal (including groundwater-surface water interaction) in combination with (mostly numerical) tools from operations research and economics (see Harou *et al* 2009, MacEwan *et al* 2017 for a more elaborate classifications).

#### *6.2.1. Economic analyses with simple aquifer models*

Most of the economic literature that is concerned with optimal groundwater use regards groundwater as temporary non-renewable resource, where it is expected that, at least for some period into the future, groundwater pumping will exceed groundwater recharge, leading to groundwater depletion and head decline. The question then is to find an optimal future trajectory of pumping rates such that the present economic value of groundwater use is maximized (e.g. Burt 1964, 1966). The pumping rates in such an optimal trajectory typically decrease over time to the groundwater recharge rate or until the aquifer is physically depleted.

Within the literature on optimal withdrawal rates over time (called inter-temporal efficiency), the following distinctions are important (Gisser and Sánchez 1980, Negri 1989): the first is controlled (inter-temporally efficient) withdrawal rates (Gisser and Sánchez 1980) versus pumping rates under full competition (Gisser and Mercado 1973). In case of controlled withdrawal, the assumption is that a single owner or all groundwater users collectively adopt a withdrawal trajectory that maximizes the present economic value of the groundwater resource. Free competition typically occurs in case there are many users of a single common resource. Under free competition, individual groundwater users cannot assume that water left in the ground is available for him/her next year, as it may be used by his/her neighbour. This stimulates individual users to forego on inter-temporal efficiency and maximize current net return by increasing withdrawal such that current marginal revenue equals marginal withdrawal costs. The second important distinction that is made pertains to the access to the groundwater resource (Negri 1989). In case of non-exclusive access, everyone that is able is allowed to use the resource. This is can be compared to fisheries without quota for instance. Exclusive access means that only a limited number of potential users are able to access the resource, with a single user or owner of an aquifer's groundwater as a limiting case. Access to groundwater is often limited to the landowners sitting on top of the aquifer, and as such is by definition exclusive. Some large aquifer systems, however, such as the Gangetic plain aquifer, underlie regions with a very larger number of small farms, with each farmer using its groundwater. Such a case could be seen as intermediate between exclusive and non-exclusive access of a common resource.

The basis for much of the work on optimal groundwater depletion is found in Hotelling (1931). This work is more relevant for absolutely non-renewable resources such as oil and mineral resources, although water is mentioned in this paper. Hotelling assumed that resources are exclusive and its volume known and introduces the inter-temporal efficiency as the extraction rate that maximizes the present value of the resource over the time period till complete depletion. An important result is that the percentage change in net-price of the resource per unit of time should equal the discount rate. This entails that the change in extraction rate per unit time is inversely proportional to the change in price per unit time, following the demand curve. He also shows that this depletion trajectory is inter-temporally efficient. Hotelling also analysed the case of a monopoly, where the optimal extraction trajectory and associated price depend on the demand function.

##### *6.2.1.1. Exclusive access and controlled withdrawal*

Burt (1964, 1966, 1967) in a series of papers laid the foundation of the theories of optimal groundwater withdrawal over time. The work has similarities with that of Hotelling (1931) for a monopolist extractor, except for the fact that groundwater is recharged and

the extraction costs are not constant, but increase with decreasing stock (increasing costs of pumping with deeper water tables or heads in the wells). The results equally apply to a single user or the total withdrawal of multiple users of a single aquifer that fully cooperate to follow the same pumping strategy. Burt used dynamic programming to find withdrawal trajectories that maximize the present value of groundwater use over time. The dynamic maximization problem in Burt's papers and later additions thereof has the following basic form:

$$\max_{q(t)} \left\{ \int_0^\infty [r(q) - c(q, h)] e^{-it} dt \right\} \tag{4a}$$

subject to:

$$\frac{dh}{dt} = h_0 + \frac{1}{An}(q(t) - R). \tag{4b}$$

The first equation denotes the present value of the profit made by exploiting the groundwater, with $q(t)$ the withdrawal rate over time [$L^{-3}$ $T^{-1}$], $r(q)$ the revenue [US\$ $T^{-1}$] that is made by using the groundwater (e.g. market value of crop yield), $c(q,h)$ the withdrawal costs [US\$ $T^{-1}$] that depend on the depth of the water table or water in the well ($h$ [L]) and $i$ is the discount rate [$T^{-1}$]. The second equation is the water balance of the aquifer with $h_0$ [L] the initial groundwater depth before withdrawal starts, $R$ the recharge rate [$L^3$ $T^{-1}$], $A$ [$L^2$] the area of the aquifer and $n$ specific yield or storage coefficient [−].

An important result from Burt (1964) is that intertemporal efficiency is achieved if, at every moment in time, the net return (revenue minus costs) from a marginal unit of extracted groundwater is equal to the present value of that marginal unit if it remains in the ground. Negri (1989) calls this marginal value the shadow price of a unit groundwater and shows that it can be considered as the present value of avoiding future pumping costs by leaving groundwater underground. In Burt (1966), the work is extended by including stochastic groundwater recharge and in Burt (1967) by increasing the withdrawal costs by the value of groundwater as a contingency stock against years with reduced recharge. Adding the contingency value yields an optimal withdrawal trajectory that results in reduced groundwater depletion. Dominico *et al* (1968) and Burt (1970) used simple aquifer models to analyse a situation where an institutional regulator allows groundwater mining up to a certain time and subsequently limits groundwater withdrawal equal to groundwater recharge. Here, the timing of stopping groundwater overuse is determined that maximizes the present value of both the non-renewable and renewable groundwater. The associated volume of non-renewable water extracted is called the optimal mining yield. Dominico *et al* (1968) also introduced the concept of 'economic limit' as the volume of non-renewable groundwater that can extracted until marginal costs exceed marginal revenue from groundwater use. Finally, Brown and Deacon (1972) extended the original analysis of Burt (1964) by considering optimal groundwater withdrawal under conditions of economic growth, the inclusion of delayed return flows, the conjunctive use of groundwater and surface water and groundwater recharge. Results show that both depletion and present value increase under economic growth. Also, optimal groundwater withdrawal rates increase with artificial recharge, and decrease when surface water is available. As expected, when farmers use both surface water and groundwater, increasing the price of surface water increases optimal groundwater withdrawal rates, leading to a larger depletion.

*6.2.1.2. Exclusive access, controlled withdrawal versus withdrawal under free competition*

Gisser and Sánchez (1980) compared the case of controlled withdrawal leading to intertemporal efficiency with free competition, where users use groundwater to maximize current profit at every moment in time. They showed that in case the storage capacity of aquifers is very large or the demand is weakly dependent on price or the costs are weakly dependent on depth, the differences in pumping rates between free competition and controlled pumping are very small. This also means that groundwater depletion and net present value are similar under both withdrawal strategies. Their analysis confirmed an empirical result obtained earlier by the same authors and was also shown to occur in case of a nonlinear demand function (Allen and Gisser 1984). This result has come to be known as the Gisser–Sánchez effect, which has led to quite a lot of debate on the generality of the results, particularly because it seems to suggest the validity of a laisser-faire strategy for groundwater use. Koundouri (2004) provided an extensive review of later studies that looked at the Gisser–Sánchez effect and showed that it seems to hold in the majority of cases except for two. The exceptions are a study by Koundouri (2000) where withdrawal costs become very large as the aquifer is close to full depletion and by Worthington *et al* (1985), where the relationship between withdrawal costs and head is very nonlinear, which occurs in e.g. an artesian aquifer, and the relationship between revenue and groundwater use is nonlinear, e.g. in case of heterogeneous land use. Similar to many of the above-mentioned analyses, Gisser and Sánchez (1980) disregarded the impact of groundwater depletion on aquatic or terrestrial ecosystems. Esteban and Albiac (2011) included this impact by extending the model of Gisser and Sánchez (1980) with the costs of damage to aquatic ecosystems (called environmental externalities). They showed that for two heavily exploited aquifers in Spain controlled withdrawal does lead to significantly higher present values and less depletion, while adding the costs of environmental externalities further decreases head decline and even leads to recovery of groundwater levels.

*6.2.1.3. Partly-exclusive (restricted) access, multiple users and controlled withdrawal*

The earlier work on optimal withdrawal pertained to a single owner or multiple owners each following exactly the same pumping strategy (full cooperation). In reality, there may be multiple landowners that not necessarily cooperate and each pursue their individual optimal withdrawal strategy. Negri (1989) used dynamic game theory to analyse two effects. The first is the number of owners sitting on top of the aquifer. His solution allowed analysing the situation between a single and many owners showing that if the number of owners increases and these owners are non-cooperative it will lead to a greater depletion of the resource and a lower net present value per land owner. In case the number of owners is very large, access becomes effectively non-exclusive which leads to complete depletion of the resource. He also analysed the effect of what is called a strategic externality, which means that groundwater users try to capture as much as possible of the groundwater, as it will otherwise be captured by the other users. This was taken into account by allowing the optimal strategy of groundwater users to change dependent how the groundwater depletion develops in time (assuming the same strategies be followed by the other users). Negri (1989) showed that this results in lower present value and larger groundwater depletion. Following Negri's work, Provencer and Burt (1993) introduced and analysed a risk externality, which arises from the fact that if precipitation and groundwater recharge is intermittent between years, a large groundwater stock safeguards against income insecurity of the groundwater users.

*6.2.2. Economic analyses using realistic groundwater representations*

The disadvantage of a bathtub-type groundwater representation is that it ignores the fact that groundwater decline is largest close to the pumping well, it does not include the effect of groundwater-surface water interaction (figure 4) and assumes aquifers to be homogeneous and to respond uniformly and instantly to groundwater pumping. These assumptions can lead to significant errors in the calculation of optimal pumping strategies, as compared with more realistic spatially explicit models (Zimmerman 1990, Brozović *et al* 2010). On the one hand, this may overstate the commonality of groundwater resources and therefore the effect of the pumping externality (Negri 1989). On the other hand, it may undervalue the effect of groundwater withdrawal on GDEs, i.e. environmental externalities (Esteban and Albiac 2011, Dumont 2013). Moreover, in cases where conjunctive use of groundwater and surface water is analysed (Brown and Deacon 1972), they are often assumed to be separate sources, while in reality they may act as communicating vessels and should be modelled as such (Pulido-Velazquez *et al* 2016).

Bredehoeft and Young (1970) were one of the first to realize this and incorporated a more realistic groundwater representation into an economic optimization framework. They proposed a synthetic but realistic case study representing a groundwater basin with recharge occurring from a stream on one part of the aquifer and a discharge area with phreatophyte vegetation in another part. Groundwater flow was modelled by a two-dimensional finite difference method. They used a numerical optimization method maximizing present value of groundwater use over a control period, while also looking at the phreatophyte water use as a proxy for ecosystem health over time. They showed that compared to no regulation, taxing groundwater use or using quotas results in reduced depletion, higher returns and higher water use of phreatophyte vegetation. Young and Bredehoeft (1972) subsequently extended this work to a real case where interaction between surface water and groundwater was included and where groundwater use has an impact on downstream surface water availability. Bredehoeft and Young (1983) showed that for this particular case, due to the intermittent nature of streamflow, it is beneficial to fully support agriculture with groundwater withdrawal as it maximizes net income and minimizes income variability (risk externality). Of course, this strategy has a significant effect on downstream streamflow.

Following the pioneering work of Bredehoeft and Young (1970) a large body of work has been published on the hydroeconomics of groundwater use, combining economic theory with realistic groundwater representations. A first extensive review of methods was made by Gorelick (1983) distinguishing between hydraulic management methods where groundwater levels and flows are primary variables to optimize and groundwater policy and allocation models where more complex optimization involving economic objectives are considered and the effects of economic policy, e.g. taxes, are evaluated. The extensive review by Harou *et al* (2009) provided a classification of modelling approaches that also pertains to realistic aquifer hydroeconomics. They distinguish between simulation and optimization. Simulation approaches evaluate the effects of scenarios and policies over time (e.g. Bredehoeft and Young 1983, Marques *et al* 2006, Steward *et al* 2009, MacEwan *et al* 2017), while optimization finds an optimal solution, which is in case of groundwater withdrawal pertains to finding economically efficient withdrawal trajectories. Optimization often uses analytical solutions from optimization methods such a calculus of variations and dynamic programming, which makes them more suitable to be used with simple aquifer parameterizations (Burt 1964, 1967, Gisser and Sanchez 1980, Negri 1989, Merrill and Guilfoos 2017), although optimization methods have also been merged with simulation methods, e.g. to simulate economic behaviour (optimal farm decisions on groundwater use and crop production) of users over time (Marques *et al* 2006), or

optimal conjunctive use of surface and groundwater (Pulido-Velazquez *et al* 2004, 2006). Also, distinction is made between holistic approaches, where hydro-economic and hydrologic/hydrogeologic models are fully integrated (e.g. Cai 2008, Pulido-Velazquez *et al* 2008, Mulligan *et al* 2014), and component-based models (Bredehoeft and Young 1983, Steward *et al* 2009, Howitt *et al* 2012, Foster *et al* 2014, Medellín-Azuara *et al* 2015, MacEwan *et al* 2017), where economic and hydrological models are coupled iteratively or where one of the components is embedded in the other in the form of a response function.

Overseeing the complete literature from simple aquifer models to holistic distributed modelling, one can observe a trend from modelling groundwater use from the perspective of intertemporal optimization, assuming full cooperation between users or a central planning organization (Burt 1964, Gisser and Sanchez 1980) to evaluating multiple, possibly non-cooperating users with game theory (Negri 1989, Provencer and Burt 1993) to using behavioural economics to simulate groups of users (Foster *et al* 2014) to agent-based modelling to include individual farmer behaviour (Steward *et al* 2009, Mulligan *et al* 2014).

The number of studies that appeared since Bredehoeft and Young (1970) is far too numerous to review completely. So, we end this section on the hydro-economics of groundwater use with realistic aquifer representations with highlighting the results of the studies that have been put forward as examples of the different approaches described above. Marques *et al* (2006) simulated conjunctive groundwater use of the Friant-Kern System in California, an area with known groundwater depletion. They show that by taxing surface water, groundwater depletion increases. Steward *et al* (2009) used an agent based model of farmer behaviour coupled to a groundwater flow model for a region in Kansas to evaluate the water savings by two policies: regulation: capping groundwater withdrawal below natural recharge in restricted areas, and by financial incentives: buying back part of the water rights. They show that both policies yield similar groundwater savings, but result in different patterns of depletion across the region. Mulligan *et al* (2014) applied a coupled groundwater hydroeconomic model to the Republican River Basin (Ogalla aquifer) comparing an optimal control regime with modelling groundwater users as agents that maximize their profits on a year-to-year basis by choosing crop types and groundwater withdrawal rates. The no-control multi-agent solution generates higher net revenue then the optimal control but at the expense of larger depletion and strong reductions in river flow (which negates the Gisser-Sánchez effect). They also show that capping groundwater to increase streamflow is not possible without severe impacts on net revenue of the region. Taxation can be effective in increasing streamflow and decreasing groundwater depletion while still creating sufficient net revenue over the area, but only if tax revenues are redistributed. This study thus showed that results from homogenous single-aquifer studies that predict that taxes may reduce depletion at similar or even higher present value (e.g. Bredehoeft and Young 1970) are not always applicable to heterogeneous aquifers and groundwater users that are myopic. Foster *et al* (2015) used a crop water model and a behavioural economic model to analyse the effects of groundwater pumping depth and well capacity on net revenue from cropping in Nebraska, showing that under falling water levels in wells, net revenue suffers much more from decreased pumping efficiency than from increased energy costs. The highly non-linear behaviour of net revenue or present value function at greater withdrawal depths makes that controlled groundwater withdrawal provides higher economic value then free competition, which is in contrast with the Gisser-Sánchez effect. Another counter-example to this effect was provided by MacEwan *et al* (2017). They calibrated a hydrological response function relating groundwater level decline to irrigation groundwater withdrawal using outputs from a hydrological model of Central Value California, and embedded it into an econometric model and compared controlled with free competition withdrawal. Looking only at pumping externalities, the difference in net revenue between controlled and free withdrawal is 5%. However, MacEwan *et al* (2017) also estimated the additional value of drought risk reserve (the ability to irrigate high-valued crops under drought) and avoided capital costs (the present value of avoiding stranded assets, i.e. dry wells, under groundwater decline). Adding these, drastically increased the added value of controlled withdrawal, which again is in contrast with Gisser-Sánchez.

The review above shows that the economic theory to understand groundwater depletion and the hydro-economic modelling tools to find economically efficient or socially optimal withdrawal trajectories are well developed. Yet, important knowledge gaps remain. For instance, as was shown in section 2.3 (figure 4), groundwater withdrawal affects surface water flow and groundwater levels with different temporal dimensions, depending on withdrawal stage. From the perspective of hydroeconomics, this leads to different types and timing of pumping externalities and opportunity costs. Thus, when deriving intertemporal efficiency including groundwater and surface water, the increase of withdrawal costs in time should take account of the intricacies of surface-water groundwater interaction. Section 5.5 shows that GDEs can be affected by groundwater withdrawal and depletion in many different ways. Despite this fact, studies on determining socially optimal withdrawal including environmental externalities (i.e. damage ecosystems) or ecosystem constraints are rare. Reasons for this may be that relating critical thresholds for ecosystem decline to actual withdrawal rates is complex, which makes it difficult to define realistic constraints. Also,

even though there exists a large body of literature on valuating ecosystem services (e.g. De Groot *et al* 2002, 2012, Costanza *et al* 2014), such valuations are difficult to translate to a pumping externality related to individual ecosystem components. The last research gap we mention follows from the fact that all hydro-economic studies relate to individual catchments or aquifer systems. As of now, no attempt has been made to provide a continental or global-scale analysis of optimal future withdrawal trajectories or expected withdrawal rates and groundwater declines under free competition. Even the less ambitious questions 'how will groundwater levels evolve under current withdrawal rates' and 'what is the depth of the economic limit (Dominico *et al* 1968) and when will it be reached?' have not been tackled. Yet, agricultural commodities produced by non-renewable groundwater are traded globally nowadays (Dalin *et al* 2017). Therefore, providing a global overview of regional groundwater overuse and associated economic sustainability of this overuse is globally important. Such analyses would require credible global groundwater models including groundwater-surface water interactions and groundwater quality (e.g. salinity), which have only recently been under development (De Graaf *et al* 2017, Reinecke *et al* 2018).

### 6.3. Economic and voluntary incentives for managing groundwater withdrawal

In order to manage and preferently reduce the use of non-renewable groundwater one can rely on three types of policy instruments (following Theesfeld 2010): (1) regulatory or command-and-control policy instruments; (2) economic policy instruments; (3) voluntary policy instruments. In countries or states where the government owns the water underground, it is possible to use quotas, the bestowal of water rights and the issuing of permits to limit groundwater withdrawal to economically or environmentally sustainable levels. In many countries, however, groundwater is individually or collectively owned by the land owners. In these cases, economic or voluntary instruments could be used.

Economic instruments to reduce or modulate groundwater use have been mentioned already in section 6.2. They can be divided into larger categories of which we mention water pricing, water markets, paying for ecosystem services and subsidies. The effects of *taxation or water pricing* have been investigated extensively (Höglund 1999, Ørum *et al* 2010, Dinar *et al* 2015). Taxation can have different effects, such as: reduce demand (Schoengold *et al* 2006, Rinaudo *et al* 2012, Mulligan *et al* 2014), decrease depletion and increase present value (Bredehoeft and Young 1970); increase investments in water savings technology (Cummings and Nercissiantz 1992, Medellín-Azuara *et al* 2015); stimulate efficient use of surface and groundwater under environmental constraints (Riegels *et al* 2013). *Water markets* have been successfully used to increase the economic efficiency of water use by allowing water to be traded between less-profitable and more profitable uses (Wheeler *et al* 2014, Xu *et al* 2018). Necessary requirements for this to work are, among others, a fixed cap on total water withdrawal (e.g. water rights), organized governance to mediate the trade and the infrastructure to move the water between users. Especially the latter is not always in place in case the main source of water is groundwater. Not often mentioned in the economic literature is *paying for ecosystem services* (Immerzeel *et al* 2008, Steward *et al* 2009, To *et al* 2012). Here, groundwater users are paid the amount of profit, preferable a bit more, that they lose by not extracting a certain volume of groundwater. Alternatively, programs for the buy-back of water rights are setup. If groundwater is a common pool resource, the challenge then is to make sure that sufficient users participate, which is more likely if net revenues drop sharply with increasing depth to groundwater (see Foster *et al* 2015). Finally, *subsidies* for investing in water savings technology are thought be of interest to reduce groundwater withdrawal. Quite a number of recent studies, however, indicate that these measures are either ineffective (Scheierling *et al* 2006) or can even increase consumptive water use (Ward and Pulido-Velazquez 2008). In irrigation, this so-called 'rebound effect' (Sivapalan *et al* 2014) is generally explained by the fact that saving more water will increase agricultural production (and thus consumptive water use through transpiration).

As is shown by the review above, economic theory suggests that if natural resources such a groundwater are collectively used without governance, it will eventually be depleted. However, based on her own field work and that of Blomquist (1987, 1988), Ostrom (1990) disproved this assumption by pointing out voluntary incentives for sustainable groundwater management. It was found that in many groundwater basins communities of users were able to self-organize themselves, without any central governance, and manage the groundwater resource without depleting it. Apparently, in time, these users had established rules among themselves that rendered the use of a common resource both economically as well as ecologically sustainable. By studying and comparing cases (e.g. in the USA, Nepal, Spain, Japan) of long-time sustainably managed common resources, Ostrom (1990) suggested eight 'design' principles that need to be part of a local self-organizing institution to sustainably govern a common resource. It is important to note that Ostrom (1990) distinguishes between using the resource (e.g. allocating withdrawal on a day-by-day basis) and provision for the resource (e.g. managing long-term groundwater stock, recharge enhancement, land use planning). Collective action for the latter is much more challenging and may require regional governance (Lopez-Gunn 2003).

Ostrom's work constitutes one of the foundations of socio-ecology, studying the interactions between humans and biophysical systems. Similarly, socio-hydrology (Sivapalan *et al* 2012, Di Baldassarre *et al* 2013) studies human-water interactions to better understand the evolvement of water resources systems under change (Montanari *et al* 2013). The rebound effect that results from subsidizing efficient water use (Ward and Pulido-Velazquez 2008) is one of the well-known examples studied in this emerging field. Socio-hydrology states that the concept of economic efficiency is not sufficient to understand the mechanisms behind non-sustainable water use, nor the effects of economic policy thereon. Inspired by related work on coupled human-natural systems modelling (Liu *et al* 2007), socio-ecological principles (Ostrom 1990) and socio-environmental modelling (Filatova *et al* 2016), socio-hydrology combines insights from the social sciences, including behavioural economics (Kahneman and Tversky 1979, Camerer *et al* 2004), with models from complexity theory, such as stylized conceptual models, game theory, machine learning approaches, and agent-based models to describe human-water interactions (Girard *et al* 2016, Giuliani *et al* 2016, Li *et al* 2017). Many of these methods have yet to be applied to the study of sustainable groundwater use, but the first papers start to appear (O'Keeffe *et al* 2018).

## 7. Conclusions and outlook

We have provided an extensive review of the state-of-the-art of research on and assessment of non-renewable groundwater resources and groundwater depletion. Global and regional studies on groundwater depletion show that physically non-sustainable withdrawal of groundwater is a global problem that is slowly becoming a ticking time-bomb for food security, while other detrimental effects, e.g. land subsidence and deterioration of groundwater-dependent ecosystems, are visible throughout the world.

As shown on many occasions before (Theis 1940, Alley *et al* 1999, Bredehoeft 2002, Konikow and Leake 2014), physically sustainable groundwater withdrawal is not necessarily equal to recharge, but involves increased capture through the intricate interplay between the groundwater and surface water system (figure 4). The critical withdrawal limit, resulting in the detachment of the groundwater level from the surface water system is important, because it moves withdrawal from an equilibrium regime, where groundwater withdrawal mostly influences streamflow and evaporation of vegetation, to a regime where it leads to persistent groundwater decline and depletion.

The estimates of current depletion rates and their evolution in time are extremely uncertain as is apparent from table 3 and figure 6. We reiterate that, because of the problem of increased capture, volume-based methods are preferable. In order to further constrain depletion estimates we need to further improve volume-based methods. Above all this requires better methods to estimate groundwater withdrawal, better groundwater models and more observations. This results in the following challenges:

1. The modelling of water demand and use from agriculture, industry and domestic sectors needs to be better constrained with reported statistics. It also needs to be improved, possibly by using advanced behavioural approaches based on machine learning and agent-based modelling.
2. We need to improve recharge estimates in global models. Most models only include diffuse recharge from soils, while especially in semi-arid regions spatially and seasonally concentrated recharge from water bodies and preferential flow may be important (Hartmann *et al* 2015). These processes need to be incorporated in global models and their importance assessed.
3. The accuracy of large-scale groundwater models hinges on the availability of information on hydrogeological properties (model parameters) of the subsoil, and observations of hydraulic head, groundwater age, groundwater salinity and streamflow (for calibration and validation). This calls for a global concerted effort of the hydrologic community to start building a database with available borehole lithology, observation and abstraction well data (quantity, quality, isotopes), hydrogeological profiles and input data of existing regional groundwater models.
4. Remote sensing of groundwater storage has been instrumental in detecting areas of groundwater decline, but these observations are still too low in resolution (GRACE) or too indirect (geodetic methods) to be of use for local to regional assessments. Further improvements in observation technology in combination with improved geophysical modelling and data-assimilation are needed to make the step to more relevant resolutions.

These four challenges are equally important in projecting future groundwater levels and assessing the physical limits to groundwater depletion, i.e. challenges 2 and 3 to answer the question 'how much groundwater is there?' and 1–4 to answer the question 'how long does it last?'. However, based on the review on the impacts of groundwater withdrawal on groundwater dependent ecosystems (section 5.5) and the hydroeconomics of groundwater use (section 6), it follows that before an aquifer is physically exhausted, it is much more likely that either environmental limits or economic limits are exceeded. Given the lack of information about global hydrogeology (challenge 3),

estimating environmental and economic limits may be less ambitious than estimating physical limits; but there are also challenges:

5. Environmental flow limits of aquatic ecosystems are poorly defined and very limited research has been done on the relationship between environmental flow limits and groundwater withdrawal. The impact of groundwater decline on wetland and dryland ecosystems is well understood, but underrepresented in global environmental studies. The importance of groundwater quality (including salinity) and temperature are acknowledged but need further research.
6. Hydroeconomic studies on economically efficient or socially optimal groundwater withdrawal trajectories should more directly consider the nature of groundwater–surface water interactions, groundwater quality (salinity) and include explicitly ecological constraints or environmental externalities. Also, a global analysis of the economic limits to groundwater withdrawal or future groundwater level decline under economically efficient withdrawal trajectories is a remaining challenge.

The final conclusion from this review is that the subjects of sustainable groundwater resources, non-renewable groundwater use and groundwater depletion have received enormous attention in the hydrological and environmental literature, as testified by the long list of publications on the subject. Despite these efforts, uncertainties about current groundwater depletion rates and the future limits of non-renewable groundwater use are still large, and considerable data and research challenges need to be overcome through concerted community efforts if we hope to reduce this uncertainty in the near future.

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## Appendix. Literature by theme, subtheme and topic

| Theme/Section | Sub-theme/subsection | Topic | Example references |
|---|---|---|---|
| 2. Some key definitions and concepts and definitions (See table 1) | 2.1 The dimensions of water use | Definition of water demand, water withdrawal, consumptive water use and return flow | Döll *et al* (2012), De Graaf *et al* (2014) |
| | 2.2 Fossil groundwater, non-renewable groundwater and groundwater depletion | Definition of fossil, young and modern groundwater; Age-depth relationship | Broers (2004), Gleeson *et al* (2015), Jasechko *et al* (2017) |
| | | Definition of non-renewable groundwater, groundwater depletion and groundwater mining | Margat *et al* (2006), Margat and Van der Gun (2013) |
| | 2.3 Groundwater-surface water interaction and the sources of pumped groundwater | Groundwater-surface water interaction under pumping; increased capture; physically sustainable and non-sustainable withdrawal | Theis (1940), Bredehoeft (1997), Sophocleous (1997), Winter *et al* (1998), Alley *et al* (1999), Bredehoeft (2002), Alley and Leake (2004) and Zhou (2009), Konikow and Leake (2014) |
| 3. Methods to assess groundwater depletion (See table 2) | 3.1 Volume-based methods | Global overview of volume-based estimates | Konikow (2011) |
| | | Using head observations to assess groundwater depletion | Scanlon *et al* (2012b), MacDonald *et al* (2016) |
| | | Using GRACE to asses large-scale groundwater depletion | Rodell *et al* (2009), Tiwari *et al* (2009), Scanlon *et al* (2016), Rodell *et al* (2018) |

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| Theme/Section | Sub-theme/subsection | Topic | Example references |
|---|---|---|---|
| | | Using groundwater models to assess regional-scale to global-scale groundwater change and depletion | Henriksen *et al* (2003), Faunt (2009), Oude Essink *et al* (2010), De Lange *et al* (2014), De Graaf *et al* (2017) |
| | 3.2 Water balance methods | Depletion estimates based on (global) hydrological models | Wada *et al* (2010), Wada *et al* (2012a, 2012b), Döll *et al* (2014), Pokhrel *et al* (2015), Hanasaki *et al* (2018) |
| | | Depletion estimates based on a combination of remotely sensed fluxes and hydrological models | Cheema *et al* (2014) |
| | 3.3 Indirect geodetic methods | Examples of observation-based studies and/or modelling of land subsidence as a result of groundwater depletion. Could be used to inversely estimation depletion. | Ortega-Guerrero *et al* (1999), Chai *et al* (2004), Motagh *et al* (2008), Galloway and Burbey (2011), Hay-Man Ng *et al* 2012, Amos *et al* (2014), Zhang and Burbey 2016, Minderhoud *et al* (2017) |
| 4. Estimates of groundwater withdrawal, groundwater depletion and groundwater storage (see tables 3–6) | 4.1 Global estimates of groundwater withdrawal | Country Statistics for groundwater withdrawal | Zektser and Everett (2004), Shah (2005), IGRAC: GGIS; https://un-igrac.org/global-groundwater-information-system-ggis; see table 3 |
| | | Data on within country variation of groundwater withdrawal | USA county level: USGS https://water.usgs.gov/watuse/ ; Mexico provincial level: CONAGUA https://gob.mx/conagua/); India provincial level: Central Ground Water Board http://cgwb.gov.in/; China provincial level: Ministry of Ecology and Environmental http://english.mee.gov.cn/ (Wada and Heinrich 2013) |
| | | Groundwater withdrawal (grid-based) from global hydrological and water resources modelling | Döll (2009), Wada *et al* (2010), Döll *et al* (2012, 2014), De Graaf *et al* (2014) , Hanasaki *et al* (2018); see table 3 |
| | 4.2 Global estimates of groundwater depletion | Estimates of global groundwater depletion from water balance based modelling | Postel (1999), Wada *et al* (2010, Wada *et al* (2012a, 2012b), Döll *et al* (2014), Pokhrel *et al* (2015), Hanasaki *et al* (2018); see table 3 |
| | | Estimates of global groundwater recharge from hydrological based modelling | Döll and Fiedler (2008), Wada *et al* (2010), Wada and Heinrich (2013), De Graaf *et al* (2014, 2017) |
| | | Estimates of global groundwater depletion using volume-based methods (incl.GRACE) | Konikow (2011), Van Dijk *et al* (2014), De Graaf *et al* (2017), Rodell *et al* (2018), see table 3 |
| | 4.3 Regional large-scale estimates of groundwater withdrawal and groundwater depletion | Collection of smaller-scale case studies on groundwater depletion | Custodio (2002) |
| | | Large-scale regional estimates of groundwater depletion | Konikow (2011), Richey *et al* (2015b), Wada (2016), see references in table 4 |
| | 4.4 Future projections of groundwater depletion | Projections of future groundwater depletion based on global hydrological models and climate and socio-econimic scenarios | Wada *et al* (2012b), Wada and Bierkens (2014), Yoshikawa *et al* (2014), see table 3 |
| | 4.5 Estimates of global groundwater volumes | Estimates of global groundwater volumes | Vernadskiy (1933), Makarenko (1966), Nace (1969), Korzun (1978), Shiklomanov (1993), L'lovich (1997), Richey *et al* (2015a), Gleeson *et al* (2015), De Graaf (2016), see table 5 |
| | | Global hydrogeological datasets | Richts *et al* 2011, Gleeson *et al* (2011), Hartmann and Moosdorf (2012), Gleeson *et al* (2014), De Graaf *et al* (2015), Pelletier *et al* 2016, Shangguan *et al* 2017, Limberger *et al* |

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| Theme/Section | Sub-theme/subsection | Topic | Example references |
|---|---|---|---|
| | | | (2018), Zamrsky *et al* (2018), see table 6 |
| 5. Impacts of groundwater withdrawal and groundwater depletion | | Overview papers about the impacts of groundwater woithdrawal and depletion | Custodio (2002), Dumont (2013) |
| | 5.1 Land subsidence | Groundwater withdrawal and land subsidence: processes | Galloway *et al* (1998), Leake and Galloway (2007, van Asselen *et al* (2009), Galloway and Burbey (2011), Minderhoud *et al* (2017) |
| | | Groundwater withdrawal and land subsidence: case studies | Ortega-Guerrero *et al* 1999, Chai *et al* (2004), Sato *et al* (2006), Motagh *et al* (2008), Galloway and Burbey (2011), Hay-Man Ng *et al* (2012), Amos *et al* (2014), Zhang and Burbey (2016), Minderhoud *et al* (2017) |
| | 5.2 Enhancement of hydrological drought | Duration of low streamflows ium-pacted by groundwater withdrawal | Tallaksen and van Lanen (2004), Wada *et al* (2013), Mukherjee *et al* (2018) |
| | 5.3 Contribution to sea-level change | Terrestrial contribution to sea-level change (including groundwater depletion) | Chao (1994), Greuell (1994), Gornitz *et al* (1994), Rodenburg (1994), Sahagian *et al* (1994a), Sahagian *et al* (1994b), Postel (1999), Gornitz (2000), Huntington (2008), Milly *et al* (2010), Church *et al* (2011), Wada *et al* (2012b), Reager *et al* (2016), Wang *et al* (2018) |
| | | Contribution of groundwater depletion to sea-level change | Wada *et al* (2010), Konikow (2011), Döll *et al* (2014), Pokhrel *et al* (2015), Wada *et al* (2016), De Graaf *et al* (2017), Hanasaki *et al* (2018); |
| | 5.4 Groundwater salinization | Shallow to intermediate depth of saline and brackish groundwater | Van Weert *et al* (2009), Margat and Van der Gun (2013) |
| | | Case studies with salinization, mainly by groundwater withdrawal | Custodio and Bruggeman (1987), Post *et al* (2018) |
| | | Analytical solutions to saline upconing (sharp interface) and critical withdrawal rates | Bear and Dagan (1964), Dagan and Bear (1968), Schmorak and Mercado (1969), Strack (1976), Reilly and Goodman 1987, Garabedian (2013) |
| | | Numerical investigations of upconing including dispersion and brackish transition zones | Reilly and Goodman (1987), Zhou *et al* (2005), Werner *et al* (2009), Jakovovic *et al* (2016), |
| | | Regional future scenario study of salinization by groundwater withdrawal | Mabrouk *et al* (2018) |
| | 5.5 Impact on groundwater-dependent ecosystems | Groundwater dependent ecosystems (general): classification, mapping and policy | Foster *et al* (2006), Eamus *et al* (2015, 2016), Rohde *et al* (2017), Doody *et al* (2017) |
| | | Aquatic ecosystems: environmental flows | Tennant (1976), Reiser *et al* (1989), Tharme (2003), Olden and Poff (2003), Smakhtin *et al* (2006), Linnansaari *et al* (2013), Pastor *et al* (2014) |
| | | Aquatic ecosystems: effect of groundwater withdrawal on environmental flows | Barlow and Leake (2012), Acreman *et al* (2014), Hendriks *et al* (2014), Gleeson and Richter (2018) |
| | | Aquatic ecosystems: effects of groundwater withdrawal on stream hydogeochemistry | Stark *et al* (1994), Caissie *et al* (1996), Sear *et al* (1999), Risley *et al* (2010), Merril and Tonjes (2014) |
| | | Terrestrial ecosystems: effects of groundwater withdrawal on groundwater depth and vegetation | Runhaar *et al* (1997), Shafroth *et al* 2000, Elmore *et al* (2006), Yin *et al* (2018) |
| | | Terrestrial ecosystems: effects of groundwater withdrawal on soil hydrogeochemistry and vegetation | Grootjans *et al* (1986), Lamers *et al* (2002), Van Loon *et al* (2009, Witte *et al* (2015) |

(Continued.)

| Theme/Section | Sub-theme/subsection | Topic | Example references |
|---|---|---|---|
| 6. Hydroeconomics of groundwater use | | Collections of papers and chapters on the economics of (non-)renewable natural resources | Halvorsen and Layton (2015), Halvorsen (2018) |
| | | Overview papers about hydroeconomic and hydroeconomic modelling | Brouwer and Hofkes (2008), Harou *et al* (2009), Bekchanov *et al* (2015), Bauer-Gottwein *et al* (2017), MacEwan *et al* (2017) |
| | 6.1 Properties of groundwater as a natural resource | Papers relating to the special properties of groundwater as a natural resource | Hardin (1968), Gisser and Sánchez (1980), Negri 1989, Moench (1992), Theesfeld (2010), Famiglietti (2014), Halvorsen and Layton (2015), Foster *et al* (2015), Müller *et al* (2017), Perrone and Jasechko (2017) |
| | 6.2 Optimal groundwater withdrawal and depletion | Simple aquifer models: Exclusive access and controlled withdrawal | Hotelling (1931), Burt (1964, 1966, 1967, 1970), Dominico *et al* (1968), Brown and Deacon (1972), Merrill and Guilfoos (2017) |
| | | Simple aquifer models: Exclusive access, controlled withdrawal versus withdrawal under free competition | Gisser and Sánchez (1980), Allen and Gisser (1984), Worthington *et al* (1985), Koundouri (2000), Koundouri (2004), Esteban and Albiac (2011) |
| | | Simple aquifer models: EPartly-exclusive (restricted) access, multiple users and controlled withdrawal | Negri (1989), Provencer and Burt (1993) |
| | | Realistic aquifer models: various economic analyses including investigating Gisser-Sanchez, environmental externalities, individual user behaviour, effects of heterogeneity, conjunctive surface water and groundwater use | Bredehoeft and Young (1970), Young and Bredehoeft (1972), Bredehoeft and Young (1983), Zimmerman (1990), Marques *et al* (2006), Cai (2008), Pulido-Velazquez *et al* (2008), Steward *et al* (2009), Brozović *et al* 2010, Howitt *et al* 2012, Foster *et al* (2014), Mulligan *et al* (2014), Foster *et al* (2015), Medellín-Azuara *et al* (2015), MacEwan *et al* (2017), Pulido-Velazquez *et al* (2016) |
| | 6.3 Economic and voluntary incentives for managing groundwater withdrawal | Economic instruments (taxing/pricing, water markers, subsidies etc) to manage water use, including groundwater withdrawal | Bredehoeft and Young (1970), Cummings and Nercissiantz, (1992), Höglund (1999), Schoengold *et al* 2006, Scheierling *et al* (2006), Immerzeel *et al* (2008), Ward and Pulido-Velazquez (2008), Steward *et al* (2009), Ørum *et al* (2010), Medellín-Azuara *et al* (2015), Rinaudo *et al* (2012), To *et al* (2012), Riegels *et al* (2013), Mulligan *et al* (2014), Wheeler *et al* (2014), Dinar *et al* (2015), Xu *et al* (2018) |
| | | Voluntary incentives to manage water use, including groundwater withdrawal; collective action | Blomquist (1987, 1988), Ostrom (1990), Lopez-Gunn (2003), Theesfeld (2010) |
| | | Papers on studying and modelling human-water interactions; sociohydrology) | Liu *et al* (2007), Sivapalan *et al* (2012), Di Baldassarre *et al* (2013), Montanari *et al* (2013), Filatova *et al* (2016), Giuliani, *et al* (2016), Girard *et al* (2016), Li *et al* (2017), O'Keeffe *et al* (2018) |

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